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Title:

Parabolic dunes and their transformations under environmental and climatic changes: toward a conceptual framework for understanding and prediction

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Abstract:

The formation and evolution of parabolic aeolian dunes depends on vegetation, and as such is particularly sensitive to changes in environmental controls (e.g., temperature, precipitation, and wind regime) as well as to human disturbances (e.g., grazing, agriculture, and recreation). Parabolic dunes can develop from the stabilisation of highly mobile barchans and transverse dunes as well as from blowouts, as a consequence of colonisation and establishment of vegetation when aeolian sand transport is reduced and/or when water stress is relieved (by increasing precipitation, for instance). Conversely, existing parabolic dunes can be activated and may be transformed into barchans and/or transverse dunes when vegetation suffers environmental or anthropogenic stresses. Predicted increases in temperature and drought severity in various regions raise concerns that dune activation and transformation may intensify, and this intensification would have far-reaching implications for environmental, social, and economic sustainability. To date, a broad examination of the development of parabolic dunes and their related transformations across a variety of climate gradients has been absent. This paper reviews existing literature, compares data on the morphology and development of parabolic dunes in a comprehensive global inventory, and scrutinises mechanisms of different dune transformations and the eco-geomorphic interactions involved. This knowledge is then integrated into a conceptual framework to facilitate understanding and prediction of potential aeolian dune transformations induced by changes in environmental controls and human activities. This conceptual framework can aid judicious land management policies for better adaptations to climatic changes.

Keywords:

Dune transformation; parabolic dune; aeolian; climatic change; eco-geomorphic response; vegetation change

1. Introduction

Desertification and associated land degradation in dry regions is responsible for increased emission and reduced sink of atmospheric carbon, currently accounting for about 4% of global emissions (Lal, 2001; Millennium Ecosystem Assessment, 2005). Land degradation and vegetation loss also result in severe reduction of global food production (Scherr and Yadav, 1996). Projections of future climatic change, in particular increases in temperature and drought severity and decreases in freshwater availability expected in various regions around the world (IPCC, 2013; Maestre et al., 2012), raise concerns that aeolian activity and desertification may be exacerbated by more active dune transformations, particularly the activation of dunes that are currently stabilised by vegetation and/or biogenic crusts (Ashkenazy et al., 2012; Forman et al., 1992; Lancaster, 1997; Le Houérou, 1996; Muhs and Maat, 1993; Muhs et al., 1996; Thomas et al., 2005; Thomas and Leason, 2005). Relatively small changes in climatological parameters may contribute to an abrupt change in vegetation cover and catastrophic shifts between states of eco-geomorphic systems (Bhiry et al., 2011; Lavee et al., 1998; Muckersie and Shepherd, 1995; Rietkerk et al., 2004; Sole, 2007; Yizhaq et al., 2007; Yizhaq et al., 2009).

Despite a growing awareness of great sensitivity of aeolian landforms to vegetation change as well as the diverse feedbacks between vegetation and sand erosion and burial, the complex eco-geomorphic interrelations between vegetation and dune landforms are not completely understood. Parabolic dunes, in particular, are common aeolian landforms that are strongly controlled by eco-geomorphic interactions. Such dunes often form where there is an adequate sand supply, unidirectional wind regime, and moderate vegetation cover (Hugenholtz et al., 2008; Hugenholtz, 2010; Lancaster, 1995; McKee and Bigarella, 1979). Under ameliorating vegetation conditions, parabolic dunes can develop from highly mobile non-vegetated dunes such as barchan dunes and transverse dunes (Hart et al., 2012; Hesp and

Walker, 2013; Reitz et al., 2010; Tsoar and Blumberg, 2002). When the vegetation cover decreases, however, parabolic dunes can be transformed back to highly mobile, non-parabolic dunes (Anton and Vincent, 1986; Hack, 1941). The development and transformations of parabolic dunes are also highly sensitive to changes in many environmental factors such as precipitation (Landsberg, 1956; Stetler and Gaylord, 1996), temperature (Wolfe and Hugenholtz, 2009), wind strength and variability (Hesp, 2002; Tsoar et al., 2009), as well as to changes in land management and other anthropogenic factors (Hesp, 2001; Tsoar and Blumberg, 2002). A brief discussion on the distribution, morphology and change of parabolic dunes was recently provided by Goudie (2011), but there has been no detailed examination of the differences in development of parabolic dunes and their related transformations on a global scale across a wide climatic gradient.

This paper reviews past research on parabolic dunes and their related transformations on a global scale, exploring mechanisms of different dune transformations and their indications in the context of climatic change, mediated by vegetation, which is then integrated into a conceptual framework of understanding and predicting dune transformations influenced by changes in environmental variables and human activities. Analysis of dune landform transformations within a conceptual framework allows for the examination of geomorphic responses to environmental fluctuations and climatic change on different temporal and spatial scales. This analysis also provides a better understanding of different dune transformation mechanisms and possible dunefield evolutions, and provides a framework for planning judicious land management practices.

2. Morphology, Development, and Migration of Parabolic Dunes

Simple parabolic dunes are U- or V-shaped dunes in plan with two trailing arms pointing upwind, a deflation basin contained within arms, and a depositional lobe at the downwind end

(Hesp and Walker, 2013; Pye and Tsoar, 1990). Vegetation, usually shrubs or trees, surrounding the parabolic dunes can resist widening of the deflation basin, whilst plants on the trailing arms can bind sand and maintain the parabolic shape of dunes. Many parabolic dunes have a slip face, and some large ones may have multiple crests and slip faces. As airflow approaches towards the dune crest, flow is compressed by the stoss slope, resulting in the increases in shear stress and sediment transport. Beyond the crest, flow expands, and may create a separation zone within which positive pressure causes reversal of flow back up the lee slope, forming 'back-eddies' (Delgado-Fernandez et al., 2013; Walker and Nickling, 2002). In the zones of flow separation, flow deceleration causes grainfall deposition which forms grainfall lamination on slip faces (Hunter, 1977). As deposition continues, avalanching occurs where the slope angle reaches the critical angle of repose. The resulting grainflow and sand flowage changes pre-existing stratification and forms cross-strata (Hugenholtz et al., 2007; Hunter, 1977).

Parabolic dunes can exhibit variable morphologies (Cooke et al., 1993; Kilibarda and Blockland, 2011) (Figure 1), governed by wind regime, sediment supply and local vegetation characteristics (Baas, 2007; Hack, 1941; Hugenholtz, 2010; Pye, 1990; Rubin and Hunter, 1987; Wasson and Hyde, 1983).

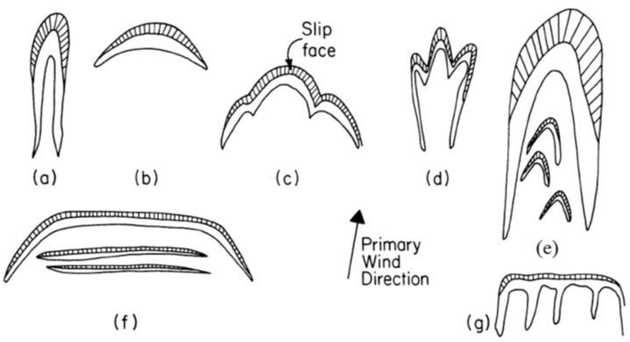

Figure 1. Diagram from Pye and Tsoar (1990) showing the following seven morphologies of parabolic dunes: (a) hairpin; (b) lunate; (c) hemicyclic; (d) digitate; (e) nested; (f) long-walled transgressive ridge with secondary transverse dunes; and (g) rake-like en-echelon dunes.

Elongated parabolic dunes with long-walled arms, also referred to as hairpin- or U-shaped dunes, develop in a strong unidirectional wind regime, whereas a greater directional variability results in much shorter trailing arms and imbricate dune forms (Gaylord and Dawson, 1987; Hesp and Walker, 2013; Pye, 1982; Pye, 1983c; Tinley, 1985). Cross-winds that blow oblique to the prevailing wind may lead to a left- or right-handed asymmetry in dune morphology. Where multiple discrete wind directions dominate at different times, hemicyclic- or digitate-shaped parabolic dunes may form (Filion and Morisset, 1983; Pye and Tsoar, 1990). The seasonal variations of winds also significantly influence airflow patterns and sediment transport over dunes (Byrne, 1997; Hansen et al., 2009).

Relatively abundant sediment supply is crucial for dunes to maintain their mobility and grow in height. The availability of external sediment supply from sandy beaches and foredunes largely controls the size of coastal dunefields (Aren et al. 2004). As dunes move forward, they can also grow in height by incorporating sand from their substrata underneath (Livingstone and Warren, 1996). If dunes move onto a non-sandy substratum in absence of an external sediment supply, the depositional lobes may flatten gradually due to continuous sand loss.

The ecological conditions and characteristics of the regional vegetation are the other essential factors determining the morphology of parabolic dunes. Digitate parabolic dunes are usually associated with the presence of a forest cover, as trees force dunes to move in divergent directions and facilitate the formation of high depositional lobes with steep windward and lee slopes (Buynevich et al., 2010; Filion and Morisset, 1983; Levin, 2011). If regeneration of the tree population is interrupted (by wildfires, for example), digitate parabolic dunes can further transform to hemicyclic dunes (Filion and Morisset, 1983). Long-lived perennial shrubs also may play an important role in trapping sediments, developing into nebkhas, and maintaining the shape of parabolic dunes (Hesp, 2008; Tsoar and Blumberg, 2002). Ephemeral annual plants, however, can only anchor dune surfaces temporarily and suffer abrupt changes under external pressures (e.g., precipitation, temperature and grazing intensities), and therefore exert minimal impacts.

Migration of parabolic dunes is principally controlled by the interplay between sand drift potential imparted by wind regime, moisture content and vegetation cover (Ash and Wasson, 1983; Bagnold, 1941; Fryberger, 1979; Lancaster, 1997; Lancaster and Baas, 1998).

The orientation of coastal parabolic dunes is largely determined by wind regime (Jennings, 1957; Landsberg, 1956), which is usually defined in terms of sand drift potential, reflecting the capacity of winds to transport sediments, as an index of regional wind energy (Arens et al., 2004; Fryberger, 1979; Levin, 2011; Levin et al., 2006; Tsoar, 2005; Wasson and Hyde, 1983).

Moisture content, which is largely controlled by precipitation and evapotranspiration, is another crucial factor for modifying sand transport and the associated dune migration (Lancaster, 1997; Tsoar, 2005). The spatial heterogeneity of soil moisture resulting from differences of local micro-environments (e.g., slope) can lead to spatial variations in sedimentation balance (Ritsema and Dekker, 1994; Stout, 2004). A greater moisture content of sand increases the critical shear velocity needed for the initiation of particle movement and inhibits sediment transport (Belly, 1964; Cornelis and Gabriels, 2003; Davidson-Arnott et al., 2008; Hugenholtz et al., 2009; Jackson and Nordstrom, 1998; Namikas and Sherman, 1995; Wiggs et al., 2004). Precipitation can, therefore, reduce the migration rate of parabolic dunes significantly (Arens et al., 2004). For example, at the Great Sand Dunes National Park and Preserve (Colorado, USA), some parabolic dunes were mobilised six times faster during drought periods than (preceding or subsequent) wet periods (Marín et al., 2005).

Unpredictable rainfall events in semi-arid regions, moreover, encourage the growth of vegetation characterised by a 'pulse-activity' response (Noy-Meir, 1973; Wand et al., 1999), and further increase surface roughness (Wolfe and Nickling, 1996). In the more humid coastal dune areas, however, precipitation variations can also significantly influence the local water table and subsequently change the sand availability, vegetation cover and dunefield mobility (Luna et al., 2012; Miot da Silva and Hesp, 2013). The spatial distribution and temporal variations of vegetation then alter the airflow dynamics over the surface, and influence the spatial heterogeneity in migration speeds of a single dune and of a dunefield as a whole (Kuriyama et al., 2005; Lancaster and Baas, 1998; Wasson and Nanninga, 1986; Wiggs et al., 1995).

Migration rates of parabolic dunes reported in literature (Table 1) naturally vary because of different local environmental settings but also depend on the measuring methods used. Typical measurement approaches, from short to long time scales, include ground surveys by setting transects and/or pins (Arens et al., 2004; Cooper, 1958; Ranwell, 1958; Wolfe and Lemmen, 1999), interpretation of multi-time aerial photographs and topographic maps (Anthonsen et al., 1996; Arens et al., 2004; Bailey and Bristow, 2004; Hesp, 2001; Hugenholtz et al., 2008; Marín et al., 2005; Pye, 1982; Siljeström and Clemente, 1990; Stetler and Gaylord, 1996; Tsoar and Blumberg, 2002), and chronological dating such as tree-ring dating and optical dating (Cooper, 1958; David et al., 1999; Wiles et al., 2003).

Table 1. Research on migration rates of parabolic dunes

Reference	Study Region	Average Migration Rate (m yr-1)	Data	Measuring Period (yrs)	Method
		Australia			
Pye, 1982	Cape Bedform, Queensland	4.8	1 dune	18	aerial photographic interpretation
	Mt. Mitchell dune, Cape Flattery, Queensland	5.6	1 dune	19	aerial photographic interpretation
Story, 1982	Northern Australia	0.05	2 dune	32	aerial photographic interpretation

Pye, 1983a	Northern Cape York Peninsula dunefield, Queensland	<6	-	-	description
Pye, 1983b	Temple Bay, Northern Cape York Peninsula	3-4	-	-	field surveys
			Brazil		
Bigarella et al., 2005	Lagoa dune field, Santa Catarina Island	2.49	1 dune	29	aerial photographic interpretation and topographic measurement of dune nose crest advance
Barbosa and Dominguez, 2004	São Francisco River Strandplain	24	-	33	aerial photographic interpretation and field measurements of the spacing between ridges
			Canada		
Hugenholtz et al., 2008	Bigstick Sand Hills, Saskatchewan	3.4	3 dunes	56	aerial photographic interpretation
Wolfe and Lemmen, 1999	Great Sand Hills, Saskatchewan	2.6	7 dunes	3	measurement of slip face advance
David et al., 1999	Seward Sand Hills, Saskatchewan	2.2	7 dunes	60	optical dating chronology and aerial photographic interpretation
			Denmark		
Anthonsen et al., 1996	Råbjerg Mile, Skagen Odde	12.2	1 dune	53	digital terrain models from topographic maps
			Israel		
Tsoar and Blumberg, 2002	South-eastern Mediterranean coast	2.8	15 dunes	46	aerial photographic interpretation
			Netherlands		
Arens et al., 2004	Kennemerland, Netherlands	3	1 dune	2	aerial photographic interpretation and erosion pin measurement
			New Zealand		
Brothers, 1954	Auckland	2.7	-	44	literature record
Hesp, 2001	Manawatu	5 - 80	-	various	aerial photographic interpretation
Muckersie and Shepherd, 1995	Manawatu	5	-	-	estimated from previous studies
Hart et al., 2012	Mason Bay	24 / 0.79	1 dune	20 / 13	aerial photographic interpretation and field surveys by a total station
			Spain		
García-Novo et al., 1976	Doñana National Park	5	-	-	aerial photographic interpretation
Arteaga et al., 2008	Liencres dune system, Cantabria	8.5	1 dune, 11 profiles	1.4	photogrammetric techniques and topographic surveys by a total station
			United Kingdom		
Ranwell, 1958	Anglesey, Wales	1.5-6.7	6 transects, 2 dunes	3	transects across dune ridges along dune moving direction
Bailey and Bristow, 2004	Anglesey, Wales	1	4 dune ridges	53	linear fit method calculated from aerial photographs
Bailey and Bristow, 2004	Anglesey, Wales	1.3	4 dune ridges	53	crest to crest method calculated from aerial photographs
			United States		
Forman et al., 2008	Cape Cod National Seashore, Massachusetts	4 / 1	12 dunes	39 / 16	aerial photographic interpretation
Marín et al., 2005	Great Sand Dunes, Colorado	7.9	13 dunes	63	remote sensing images (Landsat ETM)
Stetler and Gaylord, 1996	Hanford, Washington	1.8	-	39	stereo aerial photographic interpretation
Yurk et al., 2002	Holland, eastern shore of Lake Michigan	1.45	1 dune	61	aerial photographic interpretation

Wiles et al., 2003	Northern Chugach Mountains, Alaska	1 - 3	2 transects	200	tree-ring dating
Cooper, 1958	Oregon	1.6	-	6	measurement of slip face advance
Cooper, 1958	Oregon	<2.11	-	51	tree-ring dating
Girardi and Davis, 2010	Walking Dunes, New York	<5	1 dune	74	geo-referenced map and aerial photographic interpretation

In contrast to barchan dunes, which move forward as coherent entities, parabolic dunes continuously change in form as they elongate downwind. Whilst the arms of parabolic dunes are largely fixed in place by vegetation, the dune depositional lobes migrate at various rates. A wide morphological variety reinforces a great spatial heterogeneity in dune mobility. Precisely determining a migration rate of parabolic dunes is thus a challenge (Bailey and Bristow, 2004; Girardi and Davis, 2010). Some studies have measured the advance of slip faces (Cooper, 1958; Forman et al., 2008; Wolfe and Lemmen, 1999), whereas others have used a linear-fit method or a crest-to-crest method (Bailey and Bristow, 2004). One study, furthermore, proposed a calculus method aided with GIS technology to determine the average migration rates of lobes (Levin and Ben-Dor, 2004; Levin, 2011). Meanwhile, as a migrating rate measured is an average over a certain time period and usually also an average of a number of dunes in an area, measuring frequency and duration (in addition to the spatial scope of study sites) are also key factors for determining the migration rate.

Because of climatic instability, a dune migration rate should be evaluated on a sensible spatiotemporal scale (Lockwood, 2001). A migration rate based on a short period of field experiments (usually a few events or years) can hardly be scaled up to provide an adequate understanding of long-term dune behaviour (on a temporal scale of decades or centuries). Likewise, chronological dating elucidates long-term historical trajectories (e.g., centuries) of dunefield development, but is insufficient to assist in a detailed understanding of shorter-term variations (e.g., seasons or decades) (Aagaard et al., 2004; Sherman, 1995).

Changes in migration rates of parabolic dunes may be caused by external forces such as environmental controls and human activities (Forman et al., 2008), but also by internal or autocyclic adjustments of a geomorphological system (Brunsden, 2001). Some large parabolic dunes may migrate faster than smaller ones because of ample sand supply for wind entrainment and less vegetation impeding saltation (Marín et al., 2005). Reletively small surface roughtnesss creates less turbulence, thereby enhaning the sand transport effieicency and dune migration rate. When a dune moves onto a thicker sandy substratum, the dune migrates slower because the substratum provides more abundant sand supply to the dune. Similarly, in a sand-starving environment, as the lobe of a parabolic dune shrinks over time the dune migrates at an increasing rate. Collectively, a dune migration rate is a poor indicator of mobility of a larger dune system.

In order to estimate impacts of physical and anthropogenic variables on the development of aeolian dunes and to anticipate potential changes in dunefield mobility in the context of environmental fluctuations and climatic change, it is necessary to choose an appropriate time scale. Because dune dynamics involve time-lags and hysteresis effects between climate and dune mobility, an appropriate time scale should ensure that geomorphological components of an aeolian system have had sufficient response time to adjust themselves to external conditions such as temperature and precipitation (Hugenholtz and Wolfe, 2005; Knight et al., 2004; Levin, 2011; Yizhaq et al., 2009). The response time of dunes, however, differs depending on the characteristics of different geomorphological components (e.g., the flora) (Overpeck et al., 1992), as well as on the spatial scale (e.g., local, regional or global) of climatic change to which the dunes are responding (Huggett, 1991). Moreover, a sensible spatial scale is needed to differentiate the spatial variability of individual dune mobility from entire dunefield mobility, a variability that arises from the specific history of single dunes, for instance related to localised anthropogenic impacts.

3. Distribution of parabolic dunes

A global distribution of parabolic dunefields was collected from a comprehensive literature review of approximately 250 publications, and all sites mentioned in literature sources were examined from Google Earth imagery. Presently discernible parabolic dunefields were compiled and mapped as shown in Figure 2. Some parabolic dunefields reported in literature have been reshaped by human activities (e.g., agriculture, recreation and urbanisation), and in some regions have been largely destroyed (e.g., in Portugal, Hungary, Poland and Brazil). In China, in particular, few coastal parabolic dunefields have survived the large-scale urbanisation and industrialisation. Some other dunefields could not be identified because of insufficient image resolution and/or because they had become covered with dense vegetation.

Research on parabolic dunes started in the first half of the 20th century in the United States (Cooper, 1958; Hack, 1941; Melton, 1940), Australia (Jennings, 1957), New Zealand (Brothers, 1954) and Europe (Lefevre, 1931; Landsberg, 1956; Paul, 1944). Early research was mainly limited to qualitative description of dune morphology and local distribution, an attempt at morphology-based classification, and associated conjectures regarding dune origins and formative processes. The crucial role of wind regime on the development and morphology of parabolic dunes has been well-recognised (Bagnold, 1941; Fryberger, 1979; Jennings, 1957), but only a number of studies quantitatively measured dune morphology and migration rates (Brothers, 1954; Cooper, 1958; Ranwell, 1958).

The importance of parabolic dunes did not gain much attention until the 1980s, when research gradually expanded to a number of different regions including India (Wasson et al., 1983), Australia (Pye, 1982; Story, 1982), Fiji (Kirkpatrick and Hassall, 1981), South Africa (Eriksson et al., 1989) and Saudi Arabia (Anton and Vincent, 1986). During this period, research focused on field measurements and understandings of physical processes (flow dynamics and sand transport) and controlling factors (wind regime, sediment supply and vegetation cover). Different dune transformations have been noted in various regions, yet detailed investigation has been absent. Increasingly wide use of advanced technology such as aerial photographs, nevertheless, expanded research on a much larger spatial and longer temporal scale, which facilitated the systematic exploration and comparison of parabolic dunes in various environments and also facilitated the development of schematic models regarding dune formation and classification (David et al., 1999; Pye, 1982; Wolfe and David, 1997).

From 1995 onwards, more research on parabolic dunes has been conducted across different climatic regions both on the coast and inland. In particular, the use of computer modelling and simulation, expanded from the Werner Model (Werner, 1995), has enabled exploration of fundamental principles underlying the dynamics of dune patterns and testing of possible assumptions based on real-world observations and investigations. Parabolic dunes with trailing arms developing from blowouts have been successfully simulated by the Discrete Eco-geomorphic Aeolian Landscape (DECAL) model (Baas, 2002; Nield and Baas, 2008), whilst a dune transformation from barchan to parabolic form has been simulated with a continuous model (Duran et al., 2008). GIS technology, advances in remote sensing (e.g., LiDAR), and advances in luminescence dating techniques have further accelerated the capability and scope of investigations (Anthonsen et al., 1996; Levin and Ben-Dor, 2004; Tsoar and Blumberg, 2002; Swezey et al., 2013; Wolfe and Hugenholtz, 2009).

Presently, with progressive concerns about the potential impacts of climatic change on aeolian dune environments and associated impacts of human behaviour, parabolic dunes are receiving increased attention because of their sensitivity to changes in environmental conditions. Research on potential activation of currently stabilised parabolic dunefields imparted by climatic variations has been conducted in a few regions such as the Canadian

Prairies (Hugenholtz and Wolfe, 2005; Wolfe, 1997) and Israel (Tsoar, 2005), and will likely continue to be an important topic.

As indicated in Figure 2 and Table 2, parabolic dunes are widely spread across a large range of climatic gradients from hot equatorial savannah (Fernandez et al., 2009; Hesp et al., 2010; Hesp, 2008; Porat and Botha, 2008; Pye, 1982; Shulmeister and Lees, 1992) to warm climates (Anthonsen et al., 1996; Arens et al., 2004; Clemmensen et al., 2007; Hart et al., 2012; Morkunaite et al., 2011; Tsoar and Blumberg, 2002; Zular et al., 2013) to cold climates (Bélanger and Filion, 1991; Bhiry et al., 2011; Eyles and Meulendyk, 2012; McKee, 1966), and from humid climates (Bailey and Bristow, 2004; Bigarella et al., 2006; Levin, 2011; Ranwell, 1958; Wakes et al., 2010) to arid climates (Hack, 1941; Hugenholtz et al., 2010; Hugenholtz et al., 2008; Reitz et al., 2010; Wolfe and Lemmen, 1999; Yan et al., 2010) to hyper-arid desert climates (Anton and Vincent, 1986; Carter et al., 1990; Eriksson et al., 1989; Kar et al., 1998; Nichol and Brooke, 2011). In contrast to highly mobile barchan or transverse dunes, which may be distributed extensively across a large region, parabolic dunes are usually restricted to relatively small areas, mostly arranged in belts in coastal settings, or along inland river valleys and lake shores.

Parabolic dunes in coastal settings are strongly influenced by the geometrical alignment of the coastline relative to onshore winds (Jennings, 1957). Unidirectional onshore winds are preferable for the development of parabolic dunes, and such dunes are often associated with the initiation of blowouts on previously vegetated foredunes. Blowouts develop when vegetation cover is breached by either natural process such as increased wind erosion during periods of drought or storminess or human activity such as excessive grazing (Hesp, 2002). Sand exposed in a blowout is transported and deposited in the leeward margin, developing a bare lobe. A parabolic dune forms as the bare lobe migrates inland (cf. Section 4.3). These coastal parabolic dunes are widely seen in humid, sub-humid, and semi-arid regions, usually in conjunction with coastal foredunes in such areas as the west coast of Manawatu, New Zealand (Hesp, 2001), the Oregon coasts of the United States (Cooper, 1958), and the west and northeast coasts of Australia (Carter et al., 1990; Nichol and Brooke, 2011; Pye, 1982; Shepherd and Eliot, 1995).

Elongated parabolic dunes occur in coastal settings where relatively abundant sand supply continuously supplements sand loss from mobile lobes, without which dunes would otherwise be stabilised by vegetation. Elongated parabolic dunes are usually present in equatorial or warm climates with an ample annual precipitation yet interspersed with periodic dry seasons. These areas are generally dissipative systems and well-covered by dense forests or scrubs. Seasonal dry periods accompanied by strong onshore winds expose abundant sediments previously inundated by inter-dune lakes, which enables dune lobes to maintain mobility whereas vegetation on arms remains intact, forming long-walled arms. An alternation between wet and dry periods and strong onshore winds occurring in dry seasons are crucial in the elongation of parabolic dunes such as those on the east coasts of Australia, South America and Africa (Barbosa and Dominguez, 2004; Porat and Botha, 2008; Pye, 1982).

Digitate and hemicyclic parabolic dunes may develop on coasts that are exposed to a wind regime with multidirectional onshore winds. Presence of a woodland cover is of particular importance in forcing bare lobes moving inland in divergent directions, as can be seen on the northern coast of Brazil (Buynevich et al., 2010), and the east coasts of Fraser Island (Levin, 2011) and Queensland (Pye, 1982) in Australia.

Other important controls on the development of coastal parabolic dunes include wave power, beach morphology, storm surge, and sea level change. High wave energy dissipative and intermediate beaches have a wide and flat/gently sloping backshore. Without considerable flow disturbance, onshore winds can maintain high velocities and have a great potential for continuous landward sand transport, thereby providing abundant sediment supply for dune development (Short and Hesp, 1982). Strong storms may cause powerful wave action that removes vegetation and scars foredunes, and initiate blowouts (Hesp, 2002). The frequency and magnitude of storms, therefore, contributes significantly to shoreline destruction and coastal dune development. Sea level rise induces the near-shore profile to keep adjusting itself to a new level, which is likely to increase dynamics in sediment exchange and potentially provide greater sediment supply for aeolian sand transport (Carter, 1991; Hesp and Thom, 1990; Psuty and Silveira, 2010).

Coastal parabolic dunes are also usually found adjacent to river mouths or estuaries, where sediment from rivers provides an abundant sand supply for wind transport, as can be seen in the areas at the mouth of the Sigatoka River in Fiji (Kirkpatrick and Hassall, 1981), on the south coast of Wilderness Dune Cordons in South Africa (Illenberger and Rust, 1988), on the Oregon coast in the United States (Cooper, 1958), and on the São Francisco River Strand Plain and northeast coast of Brazil (Barbosa and Dominguez, 2004; Duran et al., 2008).

In contrast to the relatively extensive research on coastal parabolic dunes - across eighteen countries - inland parabolic dunes have been investigated in only five countries. Inland parabolic dunes are usually found in arid and semi-arid regions adjacent to river valleys and along lake margins. Their formation and development is strongly governed by regional controls: the orographic conditions and the distance to rivers or lakes.

Many of the inland parabolic dunes in western North America are derived from river sediments. In this setting, mountain ridges alter regional climate regime and the local bio-geomorphic interactions, as is the case for the widespread dunefields on the Great Plains along the eastern side of the Rocky Mountains (Forman et al., 1992; Halfen et al., 2010; Holliday, 2001; Hugenholtz, 2010; Madole, 1995; Muhs et al., 1996). The Rocky Mountains block moisture from the Pacific Ocean in the west, casting areas in the east in rain shadow (Hugenholtz et al., 2010). Most parabolic dunes in this region are stabilised under prevailing climate conditions, and LiDAR images reveal that they have been transformed from barchans under recent climate warming (Wolfe and Hugenholtz, 2009). The stabilised parabolic dunes derived from fluvial sediments are also found in the south-eastern United States such as in the Savannah River valley in Jasper County of South Carolina (Swezey et al., 2013) and on the Coastal Plain of Georgia (Ivester and Leigh, 2003).

Inland parabolic dunes can also be derived from lake sediments. The White Sands, for example, consist of gypsum sediments that precipitated as a saline lake evaporated (Kocurek et al., 2007; Langford, 2003; Scheidt et al., 2010). A small scale of parabolic dunes on the eastern shore of Lake Michigan is associated with the development of blowouts (Arbogast et al., 2002; Hansen et al., 2009; Hansen et al., 2010). On the eastern shore of Hudson Bay imbricate parabolic dunes have developed under multidirectional winds (Filion and Morisset, 1983).

In contrast to coastal parabolic dunes, which usually develop either from expansion and activation of blowouts or from stabilisation of transgressive dunefields (Hesp and Walker, 2013), inland parabolic dunes usually develop from barchan or transverse dunes. When barchans move into an environment with more abundant vegetation (e.g., closer to a river or a higher water table), their horns are invaded and anchored by grasses and shrubs first. The remaining bare lobes then move forward, leaving behind the stabilised horns (cf. Section 4.2).

Ample sand availability from mobile dunefields upwind may enable such parabolic dunes to maintain a high mobility. Examples of such parabolic dunes are widely distributed on the eastern margins of White Sands in New Mexico (McKee, 1966; Reitz et al., 2010), in the west of Fremont County and on the eastern Snake River Plain in Idaho (Chadwick and Dalke, 1965; Forman et al., 2003), and in the east of the Horqin Desert in Inner Mongolia (Yan, 2010).

Arms of inland parabolic dunes usually have relatively low relief compared with the arms of coastal parabolic dunes because grasses and shrubs rather than trees dominate these inland regions. As dune arms are frequently overridden or cut through by following dunes, elongated parabolic dunes are not commonly seen inland (Halfen et al., 2010; Marín et al., 2005), with the exception of those in the Thar Desert of India and Pakistan (Wasson et al., 1983). Although trees hardly survive in an arid desert environment, patches of stunted trees have been shown to initiate the development of small parabolic dunes in the Kalahari Desert, South Africa (Eriksson et al., 1989).

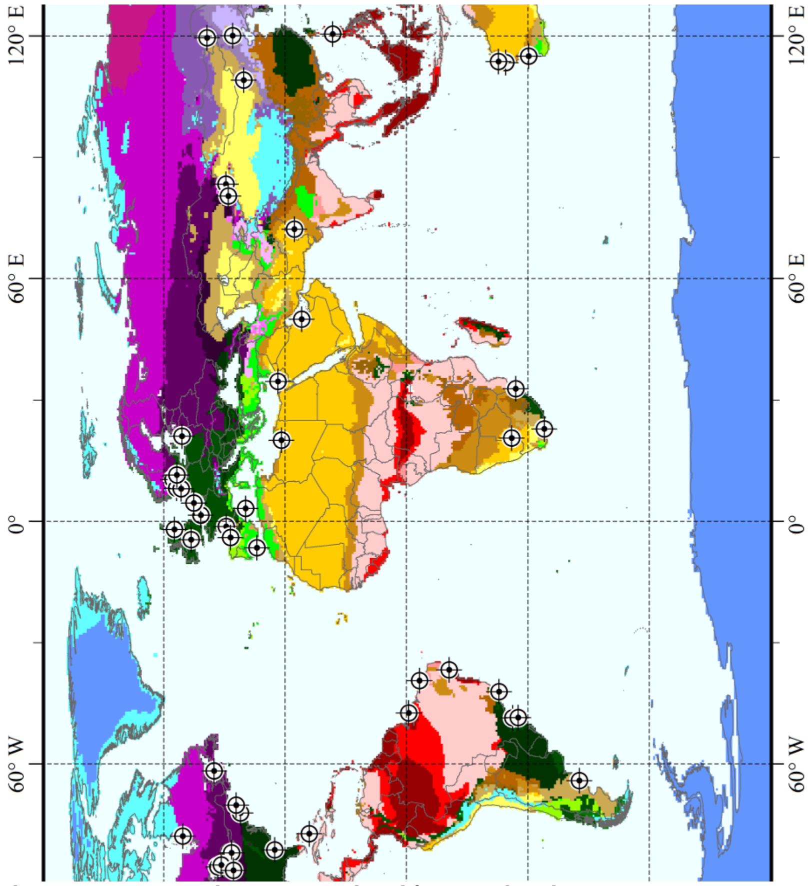

Figure 2. Global distribution of parabolic dunes. Köppen-Geiger climate zone is adapted from Kottek et al., 2006.

Table 2. Global distribution of parabolic dunes

Region	Representative Literature	Köppen-Geiger Climate Zone & Main Climate	Mean Annual Precipitation (mm)	Mean Annual Temperature (˚C)	Location
		Argentina			
Valdes Peninsula, Northeastern Patagonia	del Valle et al., 2010	BSk: arid steppe climate	steppe 231	cold arid 13	coastal
		Australia			
Cape Bedford and Cape Flattery, Queensland	Pye, 1982; 1984	Aw: equatorial savannah with dry winter	winter dry 1784	hot 27	coastal
Cervantes-Dongara coast	Shepherd and Eliot, 1995	Csa: warm temperate climate with dry summer	summer dry 534	hot summer 20	coastal
Eyre Peninsula	Dutkiewicz and Prescott, 1997	Csb: warm temperate climate with dry summer	summer dry 383	warm summer 18	coastal
Fraser Island	Ward, 2006; Levin, 2011	Cfa: warm temperate climate	fully humid 1200	hot summer 22	coastal
Groote Eylandt	Shulmeister and Lees, 1992	Aw: equatorial savannah with dry winter	winter dry 1350	hot 26	coastal
King Island, Tasmania	Jennings, 1957	Csb: warm temperate climate with dry summer	summer dry 811	warm summer 13	coastal
north of Carnarvon	Carter et al., 1990	BWh: desert climate	desert 241	hot arid 31	coastal
Northern Cape York Peninsula, Queensland	Pye, 1983b; 1984	Aw: equatorial savannah with dry winter	winter dry 1745	hot 26	coastal
Northern Territory coast	Story, 1982	Aw: equatorial savannah with dry winter	winter dry 1200	hot 27	coastal
Point Cloates, Carnarvon shelf	Nichol and Brooke, 2011	BWh: desert climate	desert 226	hot arid 22	coastal

Ramsay Bay, Hinchinbrook Island, Queensland	Pye, 1983a; 1984; Pye and Mazzullo, 1994	Am: equatorial monsoon	monsoonal 2143	hot 25	coastal
River Murray mouth region	Murray-Wallace et al., 2010	Csb: warm temperate Mediterranean climate with dry summer	summer dry 400	warm summer 16	coastal
Brazil					
Atalaia Beach, Pará State	Buynevich et al., 2010	Am: equatorial monsoon	monsoonal 2500	hot 25	coastal
Fortaleza, Ceará	Duran et al., 2008	As: equatorial savannah with dry summer	summer dry 1642	hot 27	coastal
Lagoa dune field, Santa Catarina Island	Bigarella et al., 2005; 2006	Cfa: warm temperate subtropical climate	fully humid 1521	hot summer 21	coastal
Rio de Janeiro coast	Fernandez et al., 2009	Aw: equatorial savannah with dry winter	winter dry 771	hot 24	coastal
São Francisco do Sul coastal barrier	Zular et al., 2013	Cfa: warm temperate subtropical climate	fully humid 1250	hot summer 18	coastal
São Francisco River Strand Plain	Barbosa and Dominguez, 2004	As: equatorial savannah with dry summer	summer dry 1700	hot 24	coastal
Canada					
Bigstick Sand Hills, Saskatchewan	Hugenholtz et al., 2007; Hugenholtz, 2010	BSk: continental steppe climate	steppe 380	cold arid 3	inland, river bank
eastern coast of Hudson Bay, Northern Québec	Filion and Morisset, 1983; Bélanger and Filion, 1991; Bhiry et al., 2011	Dfc: snow climate	fully humid 637	cool summer -4	coastal
Îles de la Madeleine, Quebec	Giles and McCann, 1997	Dfb: snow climate	fully humid 795	warm summer 5	coastal
Lake Huron coast, Ontario	Byrne, 1997; Eyles and Meulendyk, 2012	Dfb: snow climate	fully humid 847	warm summer 8	inland, lake shore
northern Great Plains, Saskatchewan	Wolfe and Hugenholtz, 2009	BSk: continental steppe climate	steppe 380	cold arid 3	inland, river bank
China					
Ebinur Lake district, Xinjiang	Jia et al., 2012	BWk: arid desert climate	desert 91	cold arid 8	inland, lake shore
Hobq Desert, Ordos, Inner Mongolia	Yan, 2010; Zhang et al., 2011	BSk: arid steppe climate	steppe 312	cold arid 7	inland, river bank
Horqin Desert, Inner Mongolia	Yan, 2010	BSk: arid steppe climate	steppe 360	cold arid 5	inland, river bank
Hulunbuir Grassland, Inner Mongolia	Zhuang and Hasi, 2005; Yan, 2010	Dwb: snow climate with dry winter	winter dry 354	warm summer -1	inland, river bank
Take Ermu Ku'er Desert, Xinjiang	Zeng, 2008	BSk: arid desert climate	steppe 178	cold arid 10	inland, river bank
Denmark					
Anholt, Kattegat	Clemmensen et al., 2007	Cfb: warm temperate climate	fully humid 478	warm summer 9	coastal
Lodbjerg, northwest coast of Jutland	Clemmensen et al., 2001	Cfb: warm temperate climate	fully humid 359	warm summer 9	coastal
Råbjerg Mile, Skagen Odde	Anthonsen et al., 1996	Cfb: warm temperate climate	fully humid 706	warm summer 7	coastal
Vejers, west coast of Jutland	Clemmensen et al., 1996	Cfb: warm temperate climate	fully humid 355	warm summer 9	coastal
Fiji					
Sigatoka sand dunes, Viti Levu	Kirkpatrick and Hassall, 1981	Af: equatorial rainforest	fully humid 1862	hot 30	coastal
France					
northern shore	Meurisse et al., 2005	Cfb: warm temperate climate	fully humid 592	warm summer 11	coastal
South-western coast	Bertran et al. 2011	Cfb: warm temperate climate	fully humid 823	warm summer 11	coastal
India & Pakistan					
Thar Desert	Wasson et al., 1983; Goossens et al., 1993; Kar	BWh: tropical desert climate	desert 172	hot arid 27	inland, river bank

	et al., 1998				
			Israel		
southeastern Mediterranean Coast	Tsoar and Blumberg, 2002; Ardon et al., 2009	Csa: warm temperate Mediterranean climate with dry summer	summer dry 500	hot summer 20	coastal
			Libya		
Adjabiya coast	Goudie 2011	BWh: desert climate	desert 143	hot arid 21	coastal
			Lithuania		
Curonian Spit, southeastern Baltic Sea Coast	Morkunaite et al., 2011	Cfb: warm temperate climate, intermediate between marine and continental	fully humid 660	warm summer 7	coastal
			Mexico		
El Farallon-La Mancha Dunefield	Hesp et al., 2010	Aw: equatorial savannah with dry winter	winter dry 1200	hot 24	coastal
			Netherlands		
Castricum	Jungerius and Riksen, 2010	Cfb: warm temperate climate	fully humid 847	warm summer 10	coastal
Kennemerland	Arens et al., 2004	Cfb: warm temperate climate	fully humid 847	warm summer 10	coastal
			New Zealand		
Manawatu coastal plain	Hesp, 2001; Clement et al., 2010	Cfb: warm temperate maritime climate	fully humid 900	warm summer 13	coastal
Mason Bay, Stewart Island	Wakes et al., 2010; Hart et al., 2012	Cfb: warm temperate maritime climate	fully humid 1324	warm summer 10	coastal
western coast of Auckland	Brothers, 1954	Cfb: warm temperate maritime climate	fully humid 1240	warm summer 15	coastal
			Philippines		
Ilococ Norte	Hesp, 2008	Aw: equatorial savannah with dry winter	winter dry 2067	hot 27	coastal
			Saudi Arabia		
Jafurah Desert, Eastern Province	Anton and Vincent, 1986	BWh: desert climate, Indian Ocean Monsoonal	desert 88	hot arid 27	coastal
			South Africa		
Maputaland coatal plain	Porat and Botha, 2008	Aw: equatorial savannah with dry winter	winter dry 1100	hot 22	coastal
southern Kalahari Desert	Eriksson et al., 1989	BWh: arid to semi-arid desert climate	desert 237	hot arid 18	inland, river bank
Wilderness Dune Cordons	Hellström, 1996; Illenberger, 1996	Cfb: warm temperate climate	fully humid 681	warm summer 17	coastal
			Spain		
Doñana National Park	Siljeström and Clemente, 1990	Csa: warm temperate Mediterranean climate with dry summer	summer dry 542	hot summer 17	coastal
Liencres dune system, Cantabria	Arteaga et al., 2008	Cfb: warm temperate climate	fully humid 1150	warm summer 14	coastal
Mallorca	Servera et al., 2009	Csa: warm temperate Mediterranean climate with dry summer	summer dry 427	hot summer 18	island
			United Kingdom		
Anglesey, north Wales	Bailey and Bristow, 2004	Cfb: warm temperate maritime climate	fully humid 1434	warm summer 11	coastal
Sands of Forvie, Scotland	Robertson-Rintoul, 1990; Ritchie, 2000	Cfb: warm temperate maritime climate	fully humid 750	warm summer 9	coastal
			United States		
Cape Cod National Seashore, Massachusetts	Forman et al., 2008	Cfa: warm temperate climate	fully humid 1065	hot summer 10	coastal

Casper Dunefield, Casper, Wyoming	Halfen et al., 2010	BSk: semi-arid steppe climate	steppe 300	cold arid 7	inland, river bank
Eastern Colorado	Madole, 1995	BSk: semi-arid steppe climate	steppe 380	cold arid 10	inland, river bank
eastern Upper Michigan	Loope et al., 2010	Dfb: snow continental climate	fully humid 3500	warm summer 6	inland, lake shore
Great Bend Sand Prairies, Kansas	Arbogast, 1996	Cfa: warm temperate continental climate, semi-arid to sub-humid	fully humid 678	hot summer 14	inland, river bank
Great Sand Dunes National Park and Preserve, Colorado	Marín et al., 2005; Forman et al., 2006	BSk: semi-arid steppe climate	steppe 933	cold arid 7	inland, river bank
Hanford, Washington	Stetler and Gaylord, 1996	BSk: arid steppe climate	steppe 160	cold arid 12	inland, river bank
High Plains of Colorado	Forman et al., 1992; Muhs et al., 1996	BSk: semi-arid steppe climate	steppe 912	cold arid 10	inland, river bank
Holland, eastern shore of Lake Michigan	Arbogast et al., 2002; Timmons et al., 2007; Hansen et al., 2009, 2010	Dfb: snow continental climate	fully humid 2738	warm summer 9	inland, lake shore
Lanphere Dunes, Northern California	Craig, 2000	Csb: warm temperate climate with dry summer	summer dry 969	warm summer 12	coastal
Navajo County, Arizona	Hack, 1941	BSk: steppe climate, true desert to humid mountain climate	steppe 210	cold arid 12	inland, river bank
north-western Bahamas	Kindler and Strasser, 2000	Aw: equatorial savannah with dry winter	winter dry 1120	Hot 24	coastal
Oregon coast	Cooper, 1958	Csb: warm temperate climate with dry summer	summer dry 1794	warm summer 12	coastal
Petoskey State Park, Michigan	Lepczyk and Arbogast, 2005	Dfb: snow continental climate	fully humid 813	warm summer 7	inland, lake shore
Savannah River valley, South Carolina	Swezey et al., 2013	Cfa: warm temperate climate	fully humid 1298	hot summer 18	inland, river bank
St. Anthony, Idaho	Chadwick and Dalke, 1965	BSk: semi-arid steppe climate	steppe 340	cold arid 13	inland, river bank
south Texas	Forman et al., 2009	Cfa: warm temperate subtropical climate	fully humid 806	hot summer 22	coastal
Southern High Plains of Taxas and New Mexico	Holliday, 2001	BSk: semi-arid steppe climate	steppe 409	cold arid 8	inland, river bank
Walking Dunefield, Napeague, New York	Girardi and Davis, 2010	Cfa: warm temperate climate	fully humid 1217	hot summer 11	coastal
White Sands Dunefield, New Mexico	McKee, 1966; Reitz et al., 2010	BSk: arid steppe climate	steppe 264	cold arid 16	inland, river bank
Wilderness State Park, northern lower Michigan	Lichter, 1995	Dfb: snow continental climate	fully humid 767	warm summer 5	inland, lake shore

3.1 Coastal parabolic dunes

Coastal parabolic dunes are extensively developed in Australia. Hairpin-, hemicyclic-, and digitate-shaped parabolic dunes are found on the northeast coast (Figure 3a), and some of these dunes overlap with or are nested within others, developing a compound dune form (Levin, 2011; Pye, 1982; Pye, 1983a; Pye, 1983c; Pye, 1984; Pye and Mazzullo, 1994; Shulmeister and Lees, 1992; Ward, 2006). These parabolic dunes are generally stabilised by vegetation except some in the Cape Bedford – Cape Flattery dunefields and on the east coast of Northern Cape York Peninsula. This region is controlled by an equatorial savannah climate or an equatorial monsoon climate with a strong seasonal variation of humidity in the north, to a fully humid climate down to the south. In comparison to the very elongated parabolic dunes on the east coast of Australia, parabolic dunes on the west coast (Figure 3b) are more mobile and somewhat less elongated (Carter et al., 1990; Nichol and Brooke, 2011; Shepherd and Eliot, 1995). The climate there is hot and arid, with a desert climate in the northwest, to a seasonal humid climate (dry summers) in the southeast. In the Carnarvon dunefield, for example, the mean annual precipitation is only ~200 mm. Parabolic dunes are also present on King Island (Jennings, 1957) and on the south coast of South Australia (Dutkiewicz and Prescott, 1997; Murray-Wallace et al., 2010), but on a relatively smaller scale, governed by a temperate climate with dry summers (Figure 3c).

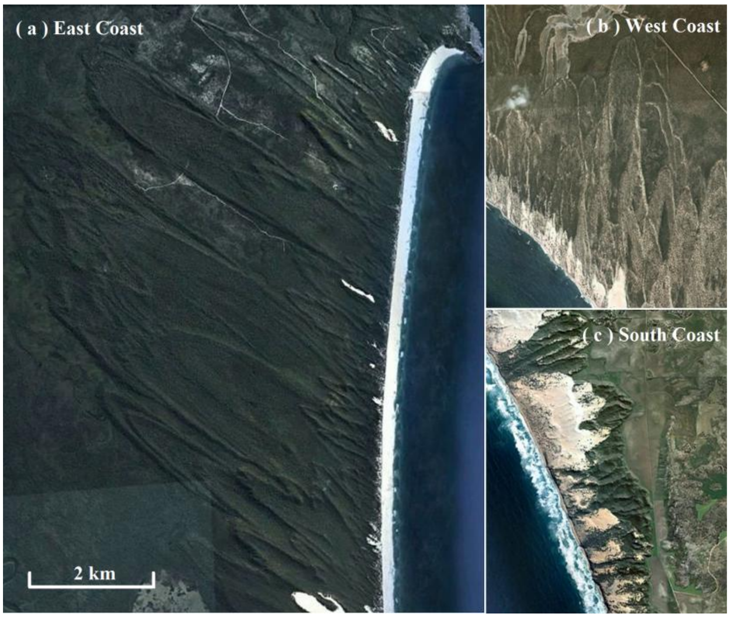

Figure 3. Coastal parabolic dunes in Australia, at the same scale.

Hairpin-shaped parabolic dunes are present along the west coast of Auckland in New Zealand (Brothers, 1954), along the west coast of Maputaland Plain and south coast of

Wilderness Dune Cordons in South Africa (Hellström, 1996; Illenberger, 1996; Porat and Botha, 2008), and at the mouth of the Sigatoka River in Fiji (Kirkpatrick and Hassall, 1981). In these coastal regions, most of the parabolic dunes are fully vegetated with minor aeolian sediment transport. In some of these locations, the movement of parabolic dunes is impeded by a forest canopy, and the dunes have developed digitate-shaped lobes. On the coasts of Mason Bay and Manawatu Plain in New Zealand, however, parabolic dunes are highly mobile and they move inland continuously (Figure 4), even though the climate is fully humid with a mean annual precipitation of ~900 mm (Clement et al., 2010; Hesp, 2001; Hart et al., 2012; Wakes et al., 2010).

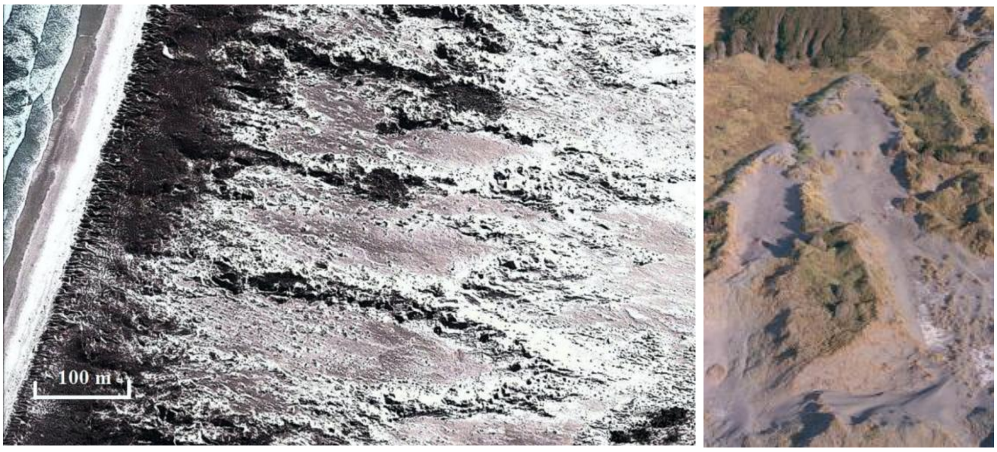

Figure 4. Coastal parabolic dunes on the lower west coast (Manawatu region) of the North Island, New Zealand (Photo courtesy of Patrick Hesp ©)

In Brazil, active parabolic dunefields are primarily concentrated on the equatorial coast (in an equatorial savannah climate with dry summers) in particular on the São Francisco River Strand Plain (Barbosa and Dominguez, 2004) and Fortaleza coast (Duran et al., 2008). In these areas, the parabolic dunes display a long-walled transgressive ridge/lobe with relatively thin and short trailing arms (Figure 5). Parabolic dunes at the São Francisco do Sul coastal barrier, however, are influenced by a humid subtropical climate and are fully stabilised by a forest canopy only with minor development of blowouts (Zular et al., 2013). Due to the impacts of human activity, few active parabolic dunes survive on the Atalaia Beach in Pará State (Buynevich et al., 2010; Miot da Silva and Hesp, 2010) and the coast of Rio de Janeiro (Fernandez et al., 2009).

Figure 5. Coastal parabolic dunes on the São Francisco River Strand Plain in Brazil

Lunate parabolic dunes are present in Oregon on the west coast of the United States (Cooper, 1958), and on the Mediterranean coast of Israel (Ardon et al., 2009; Tsoar and Blumberg, 2002). The Israeli coastal dunefield described in this literature, however, may also be interpreted as vegetated transverse dunes, and only a few have developed into somewhat parabolic shape, although this dunefield forms the basis for the parabolic dune stabilisation mechanism proposed by Tsoar and Blumberg (2002), see section 4.2. These parabolic dunes have lunate-shaped lobes yet lack well-defined trailing arms (Figure 6). A temperate climate with dry summers dominates these regions, and aeolian activity alternates with periods of stabilisation by vegetation cover. Small active parabolic dunefields also occur on the east coast of the United States, such as Cape Cod National Seashore in Massachusetts (Forman et al., 2008; Winker, 1992) and Walking Dunefield in New York (Girardi and Davis, 2010), both of which are in a temperate climate with an annual precipitation greater than 1000 mm. Presently partial-submerged parabolic dunes are also found in the north-western Bahamas (Kindler and Strasser, 2000).

Figure 6. Coastal parabolic dunes in Israel. The dune outlined in the red rectangular has been investigated as a parabolic dune in Ardon et al., (2009).

In Canada, coastal parabolic dunes are characterised by niveo-aeolian deposits that formed in a continental setting where winds transported snow and sand coincidently during the winter (Bélanger and Filion, 1991; Bhiry et al., 2011; Filion and Morisset, 1983; Giles and McCann, 1997). The cold and dry climatic conditions during winter facilitate niveo-aeolian activity compared with cool humid conditions during summer.

In Europe, coastal parabolic dunes are usually stabilised by vegetation and are relic features that formed under different climatic conditions, and the dunes are hence not easily identified because of their complex forms, full coverage by vegetation, and/or destruction by human activities. Stabilised parabolic dunes can be identified in northern Denmark

(Clemmensen et al., 2001), the Netherlands (Jungerius and Riksen, 2010), and Scotland in the United Kingdom (Ritchie, 2000; Robertson-Rintoul, 1990), whereas active parabolic dunes can be discernible on the coasts of France (Meurisse et al., 2005), Doñana National Park and Cantabrian Coast in Spain (Arteaga et al., 2008; Siljeström and Clemente, 1990), Curonian Spit in Lithuania (Morkunaite et al., 2011), Wales in the United Kingdom (Bailey and Bristow, 2004), and Vejers in Denmark (Clemmensen et al., 1996).

A large coastal parabolic dunefield is present along the edge of the Jafurah Desert in Saudi Arabia (Figure 7), and this dunefield extends from the inland area to the coast (Anton and Vincent, 1986). Although the Jafurah Desert has a mean annual precipitation less than 100 mm, the ubiquitous presence of shallow groundwater enables the survival of sparse desert scrubs, which lead to local erosion that forms parabolic dunes.

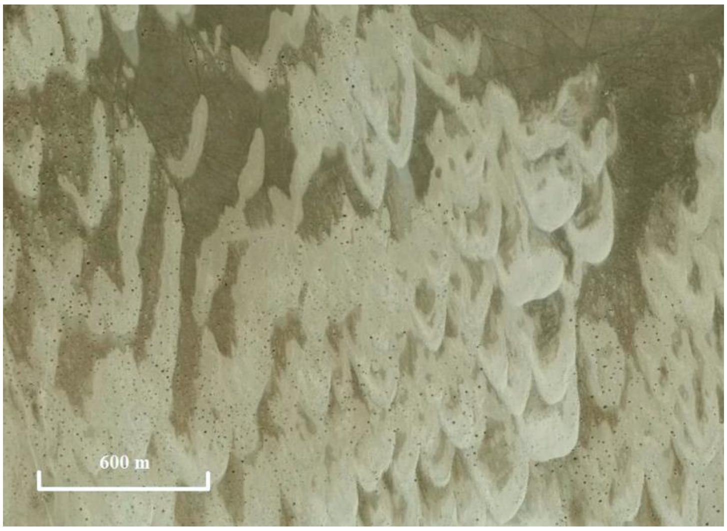

Figure 7. Coastal parabolic dunes in the Jafurah Desert, Saudi Arabia

3.2 Inland parabolic dunes

Inland parabolic dunes are located in many areas of western North America, both in the United States and Canada. On the Great Plains of western North America, large-scale aeolian sediment mobilisation is attributed to orographic impacts by the Rocky Mountains (Odynsky, 1958; Smith, 1952). Parabolic dunes are scattered widely across the Canadian Prairies, where the climate is arid continental steppe with a mean annual precipitation of ~400 mm (Hugenholtz et al., 2010; Wolfe and Hugenholtz, 2009). Most parabolic dunes here are fully stabilised by vegetation (Figure 8), although a highly active parabolic dune with development of blowouts has been studied in detail in the Bigstick Sand Hills of Saskatchewan (Hugenholtz, 2010; Hugenholtz et al., 2008; Hugenholtz et al., 2009).

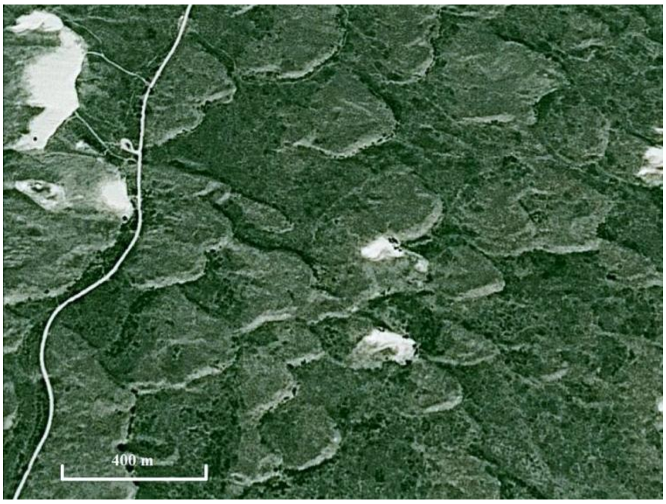

Figure 8. Inland parabolic dunes in the Canadian Prairies

Compared with those on the Canadian Prairies, parabolic dunes on the United States Prairies exhibit greater mobility. This region is governed by a semi-arid steppe climate but with varying precipitation depending on altitude and orography. Hairpin-shaped parabolic dunes occur in the Casper Dune Field in Wyoming (Figure 9) (Gaylord, 1982; Halfen et al., 2010), in the Great Sand Dunes National Park and Preserve in Colorado (Forman et al., 2006; Marín et al., 2005), and in various parts of eastern Colorado (Madole, 1995). These parabolic dunes are fully- or semi-stabilised by vegetation with trailing arms anchored by grasses and shrubs. Small active compound parabolic dunes are widespread on the High Plains of northeastern Colorado (Figure 10), Texas, and New Mexico (Forman et al., 1992; Holliday, 2001; Muhs et al., 1996), and on the Great Bend Sand Prairies in Kansas (Arbogast, 1996).

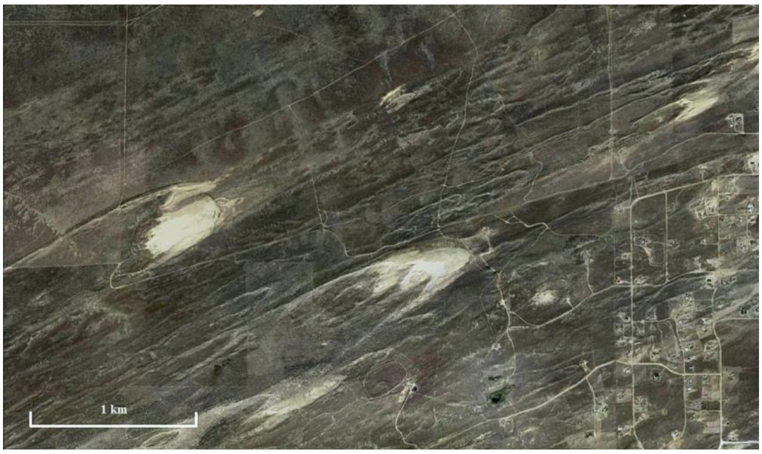

Figure 9. Inland parabolic dunes in the Casper Dune Field, Wyoming

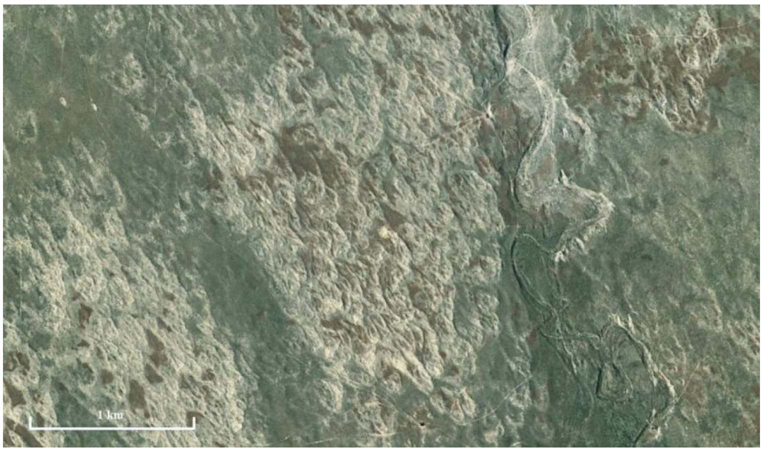

Figure 10. Inland compound parabolic dunes in north-eastern Colorado

In the United States west of the Great Plains, parabolic dunes are found on the eastern margin of White Sands in New Mexico (Figure 11) (McKee, 1966; Reitz et al., 2010), in the Navajo County in Arizona (Hack, 1941), and in the Hanford dunefield in eastern Washington (Stetler and Gaylord, 1996). These parabolic dunes, on average, have a greater mobility with evident bare lobes, compared with those on the High Plains (Figure 12).

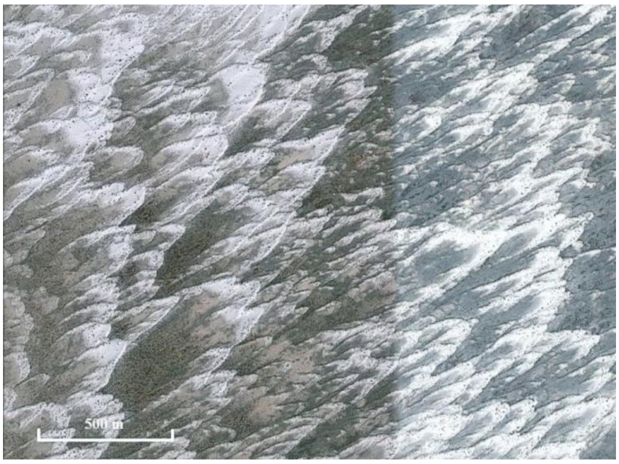

Figure 11. Inland parabolic dunes at White Sands, New Mexico

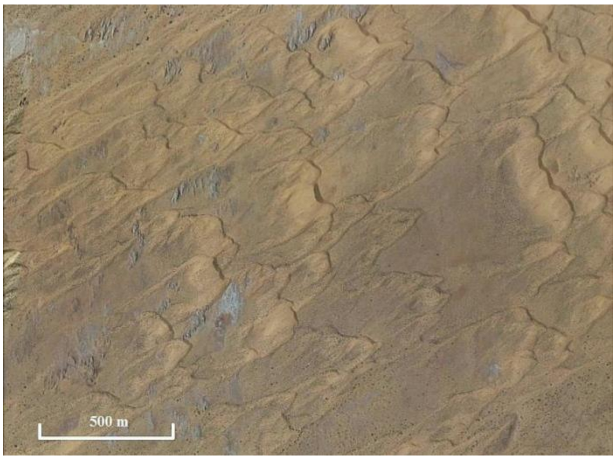

Figure 12. Inland parabolic dunes in the Navajo County, Arizona

Inland parabolic dunes are present along some of the shores of the Great Lakes of North America. For example, parabolic dunes are well-developed along the east shore of Lake Michigan and along the south shore of Lake Huron (Arbogast et al., 2002; Byrne, 1997; Eyles and Meulendyk, 2012; Hansen et al., 2009; Hansen et al., 2010; Lepczyk and Arbogast, 2005; Lichter, 1995; Timmons et al., 2007). Under the influence of a fully humid continental climate, most of these parabolic dunes are fully vegetated. Only a few that arise from blowouts are active at present.

In the Thar Desert of India and Pakistan, inland parabolic dunes link and override each other, presenting a clustered 'rake-like' appearance (Wasson et al., 1983). The irregular and asymmetric noses of parabolic dunes have developed during the process of dune overriding, in which one side of a parabolic dune is cut off when another parabolic dune moving across the nose. This process also causes the dune clusters to be irregular. The clustered parabolic dunes near Shergarh in Pakistan possess rounded noses and exhibit U-shaped dune morphology (Figure 13), whereas those between Barmer and Jaisalmer in India display V-shaped dune morphology and have exceedingly elongated arms (Figure 14).

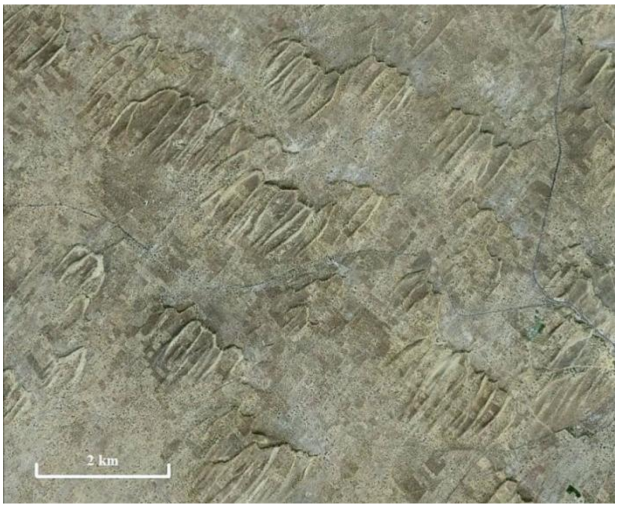

Figure 13. U-shaped parabolic dunes near Shergarh in the Thar Desert. These parabolic dunes are clustered forming individual dune groups.

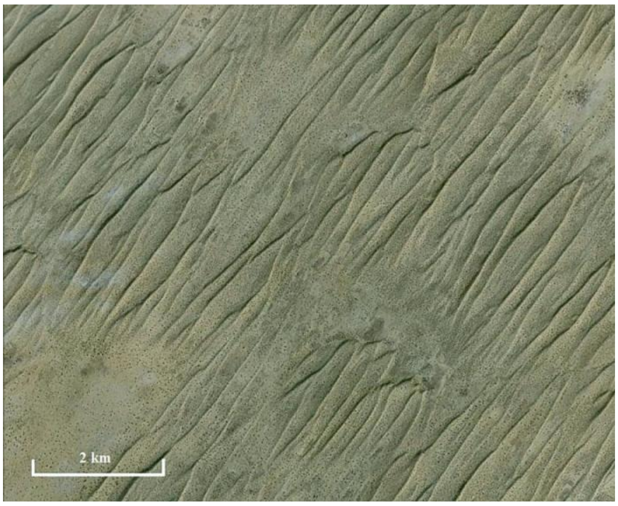

Figure 14. V-shaped parabolic dunes between Barmer and Jaisalmer in the Thar Desert. These parabolic dunes are imbricated to various degrees, and have very elongated arms.

Inland parabolic dunes in the Kalahari Desert in South Africa (Figure 15) have developed in patches under a hot and arid climate. Eriksson et al. (1989) suggested that the formation of these parabolic dunes is associated with the presence of stunted trees. Winds may continuously erode the base of the stunted trees, eventually forming blowouts and small parabolic dunes behind these blowouts.

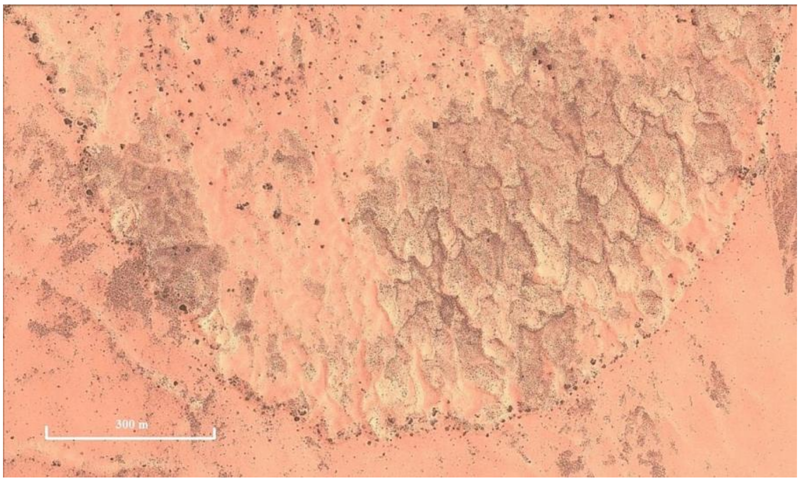

Figure 15. Inland parabolic dunes in the Kalahari Desert, South Africa

There are a variety of parabolic dunes in northern China (Yan, 2010; Yan et al., 2010). Parabolic dunes developed from blowouts are concentrated on the Hulunbuir grasslands (Zhuang and Hasi, 2005), where extensive human activity, in particular grazing and off-road driving, breaches the vegetated surface and exposes sand underneath, which initiates aeolian sand transport. The spatial arrangement of parabolic dunes in this area is hence highly controlled by local socio-economic activities, especially the related transportation networks and grazing behaviour. The Horqin Desert is located on the southeast of mountains governed by an arid steppe climate, and exhibits a transition from transverse dunes in the west to parabolic dunes in the east in similarity with White Sands in New Mexico. Highly active parabolic dunes occur extensively between the Xilamulun River and the Laoha River along the north bank of the Laoha River (Figure 16). In contrast, parabolic dunes in the Hobq Desert are fully- or partially-stabilised by shrubs (Figure 17), most of which are along the east bank of the Xuhaitu River (Yan, 2010; Zhang et al., 2011). This region is characterised by a strong seasonality of both precipitation and wind regime, resulting in dunes migrating periodically. Parabolic dunes are also found in the Ebinur Lake District (Jia et al., 2012) and the Take Ermu Ku'er Desert (Zeng, 2008) of Xinjiang.

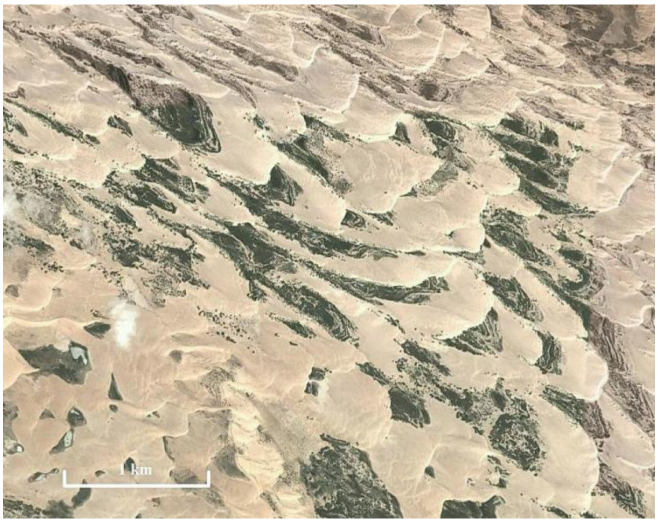

Figure 16. Inland parabolic dunes in the Horqin Desert, north-eastern China

Figure 17. Inland parabolic dunes in the Hobq Desert, northern China

4. Parabolic dune related transformations

Parabolic dunes play a significant role in dune transformations. Parabolic dunes, can not only develop from highly mobile barchan dunes, transverse dunes, and coastal transgressive dunes (McKee, 1966; Reitz et al., 2010; Stetler and Gaylord, 1996; Tsoar and Blumberg, 2002; Wolfe and Hugenholtz, 2009; Hesp and Walker, 2013), but also from blowouts (Baas and Nield, 2007; Girardi and Davis, 2010; Hesp, 2001), and stabilised transverse dunes and coastal foredunes (Carter et al., 1990; Hesp, 2002; Klijn, 1990; Muckersie and Shepherd, 1995; Nield and Baas, 2008). Well-vegetated parabolic dunes, on the other hand, can also be activated and transformed into more mobile barchan dunes and transverse dunes mediated by external pressures of both environmental changes and anthropogenic disturbances (Anton and Vincent, 1986; Hack, 1941; Hesp, 2001). In order to understand the different dune transformation mechanisms and their indications in the context of global climate change, the following sections first examine how vegetation plays a significant role in shaping aeolian dune morphology, and then explore how eco-geomorphic interactions lead to different dune transformations.

4.1 Eco-geomorphic interactions in aeolian dune environments

The development of a vegetated dunefield depends on the ability of vegetation to limit sand movement on the one hand, and on the ability of aeolian sand transport to limit the growth of vegetation on the other (Ashkenazy et al., 2012; Baas and Nield, 2010; Hack, 1941). Vegetation shapes local aeolian dune landscapes through processes of physical and biochemical interactions (Figure 18).

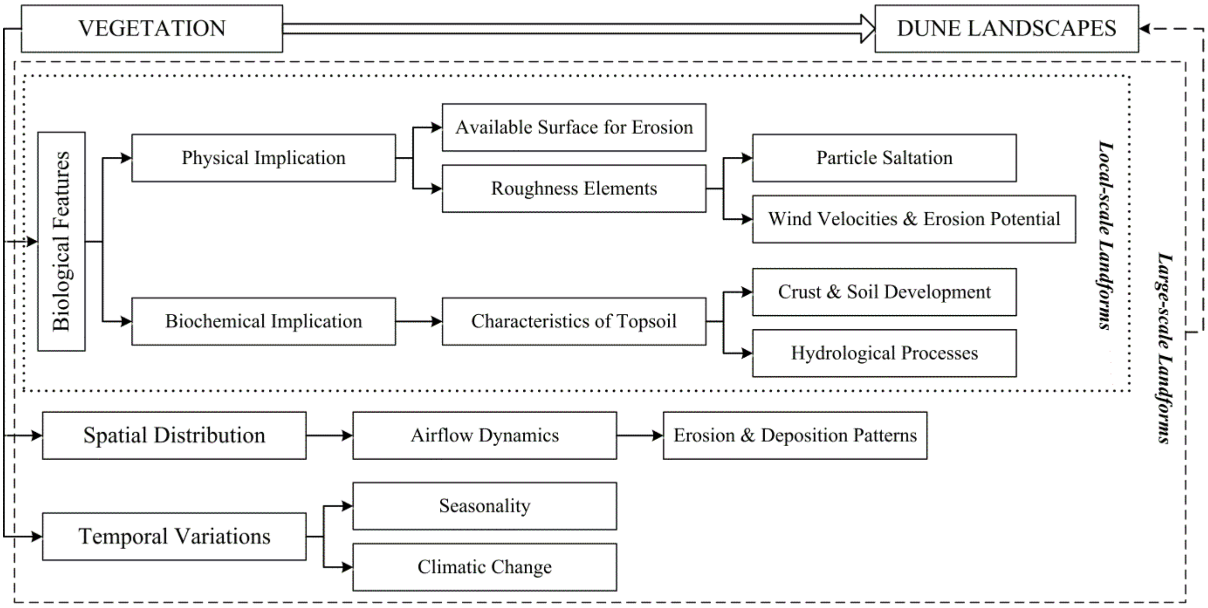

Figure 18. Schematic diagram of vegetation influences on aeolian dune landscapes

Almost all aeolian sand transport happens within 50 cm above the surface (McEwan and Willetts, 1993; Sherman et al., 1998). Physically, vegetation may shelter an erodible surface and decrease the available surface area for wind erosion. Comprising the primary roughness elements in aeolian dune environments, vegetation partitions shear stresses and hence decreases particle saltation and erosion potential because only finer grains with relatively smaller threshold shear velocities can be dislodged and transported by winds (Ash and Wasson, 1983; Gillies et al., 2007; Gillies et al., 2010; Levin et al., 2008; Musick and Gillette, 1990; Nordstrom et al., 2007; Raupach, 1992; Raupach et al., 1993; Turpin et al., 2010; Wiggs et al., 1995; Wolfe and Nickling, 1993; Wolfe and Nickling, 1996). Sand transport is effectively stopped when vegetation ground cover reaches ~20% (Kuriyama et al., 2005; Lancaster and Baas, 1998; Wiggs et al., 1995), although it is recognised that the threshold very much depends on the plant species, plant geometry and structure, and spatial distribution (Buckley, 1987). The roughness density or lateral cover, defined as the ratio of the total frontal-silhouette areas (perpendicular face of a plant which is vertical to the wind direction) of roughness elements to the total surface area (Marshall, 1971), is commonly used to evaluate the degree to which vegetation protects surfaces against wind erosion (Burri et al., 2011; Musick and Gillette, 1990; Raupach, 1992; Raupach, 1994; Raupach et al., 1993).

Whilst vegetation limits the capacity of winds to transport sand, vegetation considerably promotes its potential for trapping blown sand. This process can have significant biochemical implications for the micro-environment. By continuously trapping finer grains carried by winds, plants alter the particle-size distribution of sediments in their vicinity, in addition to altering soil texture and structure (Jungerius et al., 1995; Shields and Drouet, 1962). A relatively stable surface, a better water-retaining soil structure, and greater nutrient content (from plant roots and litter) promote the formation and development of crusts underlying plants (Johansen, 1993). Biological soil crusts (usually arising from cyanobacteria, mosses, and lichens) further increase soil stability and resistance to wind erosion by binding surface particles, and contribute nutrients to plants by fixing atmospheric nitrogen (Abed et al., 2013; Belnap, 2002; Belnap and Gillette, 1997; Belnap and Gillette, 1998; Delgado-Baquerizo et al., 2013; Drahorad et al., 2013; Eldridge and Leys, 2003; Johnson et al., 2007; Pluis, 1994; Shields et al., 1957; Thiet et al., 2005). Biological soil crusts, moreover, can resist long periods of droughts and desiccation, and can recover biological activity quickly as long as sufficient water is available from dew or precipitation (Veste et al., 2001; West, 1990). The formation of Calcretes and gypcretes can also play an important role in reducing aeolian sediment transport and stabilising mobile dunes (Amit, 1995; Chen, 1997; Dijkmans et al., 1986; Galloway et al., 1992; Pye, 1980; Swezey, 2003; Warren, 1983). Aeolian dunes in southern Tunisia, for example, are stabilised by 0.1 to 0.5 m thick gypcretes (Swezey, 2003).

Changes in the characteristics of topsoil significantly alter hydrological regimes of aeolian dune environments, in particular water distribution and budget. In aeolian dune environments, precipitation can easily be lost due to the low moisture tension and the high hydraulic conductivity pertaining to sand (Tsoar and Blumberg, 2002). Plant canopies and plant litter increase interception, lower soil bulk density, and act as a buffer protecting water from rapid loss (West, 1990). Biological crusts also influence infiltration, percolation, moisture retention, overland flow, and water redistribution (Belnap, 2006; Chamizo et al., 2012; Johansen, 1993; Rodríguez-Caballero et al., 2013; Tsoar and Moller, 1986; Verrecchia et al., 1995; West, 1990; Yair, 1990).

The complex mosaic pattern of vegetation in aeolian dune environments is responsible for spatial differences in flow dynamics over the surface and the associated patterns of sediment erosion and deposition (Ranwell, 1958; Willis and Yemm, 1961). Plants act as obstacles that cause sand to accumulate in their vicinity, thereby modifying the topography of dunes that they inhabit (Raupach, 1992; Wolfe and Nickling, 1993). Resulting sand accretion in and behind plants can lead to the formation of nebkhas (Tengberg, 1995; Tengberg and Chen, 1998) and shadow dunes (Gunatilaka, 1989; Hesp, 1981), respectively. Changes in micro-topography further alter airflow patterns over surfaces and shape dune landforms at a larger spatial scale (Frank and Kocurek, 1996a; Frank and Kocurek, 1996b). Nevertheless, vegetation, sand transport, and dune development are interactive within the context of climatological background, which is subject to a variety of environmental fluctuations because of seasonal or year-to-year variations in wind regime, temperature, precipitation, water table, and salinity. During drought or windy periods, for example, intense aeolian sediment transport can reshape dune landscapes significantly (Anderson and Walker, 2006; Byrne, 1997).

In order to survive in aeolian dune environments, plants employ both avoidance and tolerance strategies to cope with environmental stresses such as high wind velocities, sand blasting, sand accretion, wind erosion, unstable substratum, high soil temperature, and nutrient deficiency (Hesp, 1991; Maun, 1994; Maun, 1998). Seedling recruitment, for instance, usually happens during periods of low wind energy and high moisture availability

(Maun, 1994). Survival on an eroding surface is usually a challenge for most plants. They are likely die of desiccation when their roots are exposed to the air by constant erosion (Lee and Ignaciuk, 1985; Maun, 1981).

Many plants, however, have various capabilities of withstanding sand burial. According to the tolerance to sand accretion, Maun (1998) classified plant species in aeolian dune communities into the following three categories: non-tolerant, sand-tolerant, and sand-dependent. In his model (Figure 19), an individual plant may show the following different responses as sand burial increases: I) a negative response that causes the plant to die soon; II) no response and the plant grows normally within a certain level of sand accretion; and III) a stimulation of plant growth within a certain level of sand accretion. Despite a broad spectrum of the maximum tolerance to sand burial, every plant species will show a negative response beyond a certain limit (Dech and Maun, 2005; Gilbert and Ripley, 2010; Ievinsh, 2006; Kent et al., 2001; Maun, 1998; Maun and Lapierre, 1984; Wagner, 1964; Zhang and Maun, 1994).

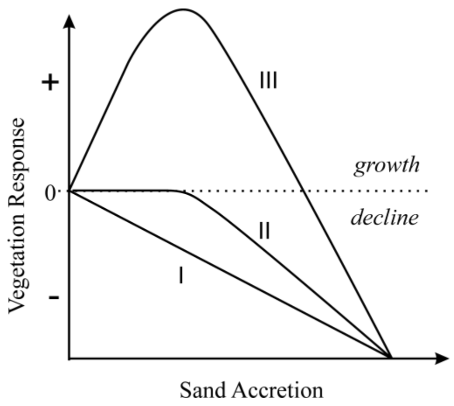

Figure 19. A model of interaction between vegetation response and increasing levels of sand accretion in aeolian dune environments (adapted from Maun, 1998). I. negative vegetation response; II. no vegetation response within a limited level of sand accretion; and III. a stimulation of vegetation growth within a certain level of sand accretion.

Burial acts as a strong selective force by eliminating sensitive plant species, decreasing the abundance of less tolerant species, and increasing the abundance of sand-tolerant and sand-dependent species (Dech and Maun, 2005; Eldred and Maun, 1982;

Martinez et al., 2001; Maun, 1994; Maun, 1998; Maun and Perumal, 1999; Moreno-Casasola, 1986). Many studies have indicated that some plant species are well-adapted to dynamic and recurrent burial, and that some plant species even require a certain amount of regular burial in order to maintain their high vigour (Bendali et al., 1990; Harris and Davy, 1987; Maun, 1998; Olson, 1958).

4.2 Transformations from barchan/transverse dunes to parabolic dunes

In contrast to parabolic dunes with two trailing arms pointing upwind, typical barchan dunes are crescent-shaped with two horns extending downwind and a slip face in their interiors. Barchan dunes develop under conditions of a high-energy unidirectional wind, a low sediment supply and a sparse vegetation cover (Hack, 1941; Rubin and Hunter, 1987; Wasson and Hyde, 1983). As sand supply increases, barchan dunes merge laterally into crescentic dunes with continuously sinuous ridges (Lancaster, 1995; Reitz et al., 2010). If sand supply is ample, then transverse dunes emerge with taller and continuous slip faces.

An increase in vegetation cover may lead to a transformation from barchan dunes or transverse dunes to parabolic dunes. This phenomenon has been studied in many countries including Israel (Tsoar and Blumberg, 2002), Brazil (Duran et al., 2008), Denmark (Anthonsen et al., 1996; Landsberg, 1956), the United Kingdom (Landsberg, 1956), New Zealand (Hesp, 2001), Canada (Wolfe and Hugenholtz, 2009) and the United States (Hack, 1941; McKee, 1966; Stetler and Gaylord, 1996) (Table 3). Most of these studies attribute the transformation to a progressive increase in vegetation cover caused by either climatic change or human disturbances. In many of these studies, however, the exact transformation process and the underlying mechanism remain unclear.

Table 3. Studies on transformations from barchan or transverse dunes to parabolic dunes

Reference	Study Region	Method	Transformation Cause
Hack, 1941	Navajo Country, Arizona, US	field surveys and photo-engravings	aggressive plants survive sand burial
Landsberg, 1956	UK and Denmark	field observation	wetter climate
McKee, 1966	White Sands, New Mexico, US	analysis of cross-stratification	arms anchored by vegetation
Muckersie and Shepherd, 1995	New Zealand	radiocarbon dating	El Nino-Southern Oscillation
Anthonsen et al., 1996	Rabjerg Mile, Skagen Odde, Denmark	topographical maps and aerial photographic interpretation	climate changes (wind regime and vegetation cover)
Stetler and Gaylord, 1996	Hanford, Washington, US	A regional climate model	increased precipitation
Hesp, 2001	Manawatu, New Zealand	aerial photographic interpretation	human activity (vegetated by farmers and less grazing pressure)
Tsoar and Blumberg, 2002	Southeastern Mediterranean Coast, Israel	aerial photographic interpretation	decrease of human pressure (stop of agricultural land-use and reduced grazing activity)
Duran and Herrmann, 2006	-	continuous modelling	the ratio between dune characteristic erosion rate and vegetation growth velocity
Duran et al., 2008	Ceara, Brazil	field surveys, QuickBird panchromatic satellite images interpretation and continuous modelling	vegetation growth rate defined by two parameters: the initial vegetation growth rate when no sand transport occurs and the maximum height a plant can reach
Ardon et al., 2009	southern coastal plain of Israel	field surveys and aerial photographic interpretation	land use change and emergence of shrubs on the crest
Wolfe and Hugenholtz, 2009	Northern Great Plains, Canada	LIDAR imagery interpretation and optical stimulation luminescence dating	climate warming
Reitz et al., 2010	White Sands, New Mexico, US	LIDAR imagery and aerial photographic interpretation	dune surface erosion/deposition rates decrease below a threshold of half the vegetation growth rate
Barchyn and Hugenholtz, 2012a	-	CA modelling	climate shift
Barchyn and Hugenholtz, 2012b	-	process-based hypothesis	the slipface deposition rate < the peak deposition tolerance of vegetation
Hart et al., 2012	Mason Bay, New Zealand	field surveys and aerial photographic interpretation	marram grass invasion

Some studies have suggested that the transformation is rooted in the anchor-like function of vegetation on the horns of barchan dunes (Livingstone and Warren, 1996; Muckersie and Shepherd, 1995; Reitz et al., 2010; Robertson-Rintoul, 1990; Stetler and Gaylord, 1996; Wolfe and Hugenholtz, 2009). Vegetation retards movement of horns, acting as anchors, whilst the dune apex continues to migrate downwind. As vegetation cover extends from dune horns to the apex, the advancing apex leaves behind the protected trailing ridges, and the dune is gradually transformed from the barchan to the parabolic shape in plan. Duran and Hermann (2006) simulated this process using a continuum model starting with a single barchan dune on a non-erodible bed. Barchyn and Hugenholtz (2012a) simulated a barchan-to-parabolic dunefield transformation imposed by a climate shift using a cellular automaton (CA) model (Baas and Nield, 2010; Nield and Baas, 2008). A few studies have suggested that this transformation happens when the sand deposition rate decreases below a certain threshold related to the vegetation growth rate (Barchyn and Hugenholtz, 2012b; Duran and Herrmann, 2006; Reitz et al., 2010).

The ‘horns-anchoring’ mechanism described above is likely to happen on the following three conditions: 1) barchan dunes are surrounded by well-vegetated land; 2) vegetation species are sufficiently aggressive to withstand a certain amount of sand burial; and 3) sediment availability is relatively limited. The first prerequisite can be fulfilled when a barchan dune is moving onto an area with a greater vegetation cover (Reitz et al., 2010). As an alternative, the interdune areas of barchan dunefields may be re-vegetated because of changes in environmental factors, such as increased precipitation (Landsberg, 1956; Stetler and Gaylord, 1996), reduced wind strength (Anthonsen et al., 1996), and climate warming (Wolfe and Hugenholtz, 2009), or because of changes in anthropogenic pressure on the environment, such as reduced grazing activity or artificial vegetation restoration (Hesp, 2001). The second condition depends on the characteristics of local plant communities, in particular the dominant species, which is also closely related to the regional climate parameters of wind regime, temperature, and precipitation. The third condition is determined by sand sources, specifically a limited external sand supply and a thin sandy substratum.

Tsoar and Blumberg (2002) proposed another potential mechanism driving the transformation from barchan dunes or transverse dunes to parabolic dunes, specifying that the establishment of vegetation on the crest of barchan dunes or transverse dunes initiates the transformation. Their argument is that vegetation preferentially grows and recovers on the crest of dunes where the erosion/deposition balance is neutral. The establishment of vegetation on the crest then changes the airflow and sediment transport dynamics over the stoss slope. Sand eroded from the stoss slope is partly trapped by clumps of plants forming isolated nebkhas, and the associated abrupt reduction in sediment supply encourages plants to take root on the lee slopes and slip faces. The accumulation of sand on the crest by nebkhas, meanwhile, gradually changes the profile of the stoss slope from convex to concave. The subsequent funnelling effect of the wind over the concave stoss slope can undercut the nebkhas and expose plant roots on the central apex, but plants on dune sides remain intact and are left behind by the mobile dune apex, developing into trailing arms eventually. Hugenholtz et al. (2008) confirmed the important role of vegetation on the dune crest in trapping sand transported from the stoss slope and in changing the dune profile.

In contrast to the previous mechanism in which vegetation is established close to the groundwater table in interdune areas and on the horns of barchan dunes, in the mechanism proposed by Tsoar and Blumberg (2002) vegetation starts to germinate and grow on the crest of dunes where erosion and deposition balance each other. Some studies have shown that a small amount of sand burial may prevent atmospheric desiccation, increase relative humidity around seeds, and anchor seedlings into soil; therefore, a small amount of sand burial may be vital to ensure successful seed germination and plant growth (Maun, 1998). Maun and Lapierre (1986) suggested that a maximum germination rate occurs at a burial depth of 2-4 cm across all four studied dune species. A study on seven dune species by Zhang and Maun (1994) has shown that the majority of seeds germinate at a depth of 5-10 cm, and that deeper burial greater than 15 cm significantly inhibits seed germination. These results are consistent with the findings from Lee and Ignaciuk (1985). A relatively stable surface with a slight amount of net deposition is clearly crucial for seeds to take root.

This ‘nebkhas-initiation’ mechanism proposed by Tsoar and Blumberg (2002) involves the following steps/prerequisites: 1) seeds germinate successfully on dune crests during a less windy season; 2) sufficient precipitation and minor aeolian sediment mobilisation allow seedlings on the crests to thrive such that they can prepare themselves adequately for strong wind energy during the following windy season; and 3) plants develop tap roots that allow them to use relatively sustainable groundwater during a dry and windy season, ensuring that the plants are able to keep vitality, trap sand continually, and develop into nebkhas that can subsequently alter the profile of stoss slopes in shape. These three prerequisites are intimately connected with each other and fundamentally provide an opportunity for vegetation to survive on the crest of dunes, arising from some combination of increased precipitation, reduced windiness, and/or rise of the local groundwater table. These variables, in turn, respond to seasonal fluctuations, climate change, and/or human pressure. To act as an initial trap and further develop into a nebkha that initiates this transformation process, the vegetation involved should be perennial and needs not only to grow up in height quickly, but also to display a branching growth pattern in order to achieve a high sand-trapping efficiency (Livingstone and Warren, 1996). Isolated plants that have lost their lower branches and leaves are much less efficient in decreasing wind velocity and sand transport, because almost all sand transport happens close to the ground. Extensive roots are necessary for vegetation survival on dune crests where soils have particularly low fertility and low capacity for retaining precipitation.

The interaction between vegetation and sand transport as well as the related mechanisms of the transformation from barchan/transverse dunes to parabolic dunes is likely to vary depending on vegetation species and the limiting factors that control vegetation germination and growth. Annual grasses such as *Agriophyllum squarrosum* are short-lived with shallow roots, are unlikely to access groundwater, and hence only survive on precipitation. They usually germinate and grow rapidly after episodic rainfall events, but die of drought shortly afterwards. The impacts of annuals on dune morphology are, therefore, generally highly limited. Perennial vegetation, however, develops deep and/or widespread roots, and can exert different impacts depending on their growth forms. For perennial grasses that grow uniformly over the surface, the degree to which such grasses reduce sand transport is primarily determined by an overall coverage of grass assembly. Perennial vegetation in need of a large water supply is usually distributed relatively sparsely as discrete clumps and shrubs. Such perennial vegetation influences the local wind regime more as individual entities through their outstanding canopies.

The 'horns-anchoring' transformation mechanism tends to occur in an environment where water deficiency is the limiting factor for vegetation growth, whereas the 'nebkhas-initiation' mechanism tends to occur where wind erosion is the predominant limiting factor. In comparison with the 'horns-anchoring' mechanism in which any perennial species can play the anchoring role as long as it can withstand a small degree of wind erosion and sand burial (Reitz et al., 2010; Wolfe and Hugenholtz, 2009), perennial vegetation in the 'nebkhas-initiation' mechanism needs to develop tap roots in order to access groundwater from dune crests and at the same time has the capability of withstanding substantial amounts of sand burial as the nebkha form (Tsoar and Blumberg, 2002). The 'nebkhas-initiation' mechanism demands more specialised plant species and is hence less common. Parabolic dunes formed by the 'nebkhas-initiation' mechanism, furthermore, are much less elongated compared with that of the 'horns-anchoring' mechanism in which bare lobes can move forward unimpeded over a long period.

4.3 Transformations from blowouts to parabolic dunes

The evolution of blowouts can result in the formation of parabolic dunes, specifically on vegetated sandy surfaces (Gutierrezelorza et al., 2005; Landsberg, 1956; Pye, 1982). Coastal examples have been documented in Australia (Pye, 1982; Pye, 1983b), New Zealand (Brothers, 1954; Hesp, 2001), the United Kingdom (Ranwell, 1958), the Netherlands (Klijn, 1990), Spain (Gutierrezelorza et al., 2005), Brazil (Duran et al., 2008), and the United States (Girardi and Davis, 2010; Hansen et al., 2009), whereas inland examples have been documented in the southern Kalahari Desert in south Africa (Eriksson et al., 1989) and the Jafurah Desert in Saudi Arabia (Anton and Vincent, 1986) (Table 4). Some early studies refer to parabolic dunes developed from blowouts as “blowout dunes” (Brothers, 1954; Cooper, 1958; Eriksson et al., 1989; Melton, 1940).

Blowouts are saucer-, bowl-, cup-, or trough-shaped depressions that usually develop by wind erosion on a pre-existing sand deposit (Hesp et al., 2011; Hesp, 2002; Hesp and Hyde, 1996). The initiation of blowouts may be associated with both natural and anthropogenic disturbances to a vegetation cover such as wildfires, increased windiness, rises in sea level, increased frequency of drought, large volcanic eruptions, storms, overgrazing, trampling, and diseases (Gutierrezelorza et al., 2005; Muckersie and Shepherd, 1995; Pye and Tsoar, 1990; Watt, 1937). These agents locally exterminate vegetation, breach the surface crust, and create discrete bare patches that develop into hollows as erosion continues. These hollows may become enlarged, and turbulent eddies may form, which further accelerate expansion of the hollows (Hesp, 2002). The deepening of the hollows also encourages the funnelling of winds, which in turn can result in more intense erosion (Smyth et al., 2014). Continued wind scour widens blowouts by steepening side walls and inducing side wall failure (Carter et al., 1990). Sand eroded from the deflation hollow is transported by winds and accumulates in the leeward margin, forming a depositional lobe, which is regarded as part of a blowout by some researchers (Hesp and Hyde, 1996; Robertson-Rintoul, 1990).

Cooper (1958) defined two primary types of blowouts: saucer blowouts and trough blowouts. Despite a wide range of variability in different aeolian environments, most blowouts can be classified as either of these two types (Hesp, 2002). Saucer blowouts are shallow, and semicircular- or saucer-shaped. Trough blowouts are deeper, and more elongated with longer lateral walls (Carter et al., 1990; Cooper, 1958; Hesp, 2002; Hesp and Hyde, 1996). Compared with a short, wide, radial depositional lobe found in the leeward margin of a saucer blowout, a trough blowout develops a tall, parabolic-shaped depositional lobe. In a trough blowout, corkscrew airflows may develop in the deflation basin. These airflows are then compressed and accelerated towards the crest of the depositional lobe, and subsequently decelerate rapidly due to flow expansion upon the exiting trough, leading to the development of a parabolic lobe (Hesp, 2002).

If a trough blowout and its lobe overcome the surrounding vegetation, they continue to migrate and develop downwind, enlarging overtime and evolving into a fully developed parabolic dune (Carter et al., 1990; Hesp and Hyde, 1996; Livingstone and Warren, 1996). The parabolic dune may continue to increase in size if the blowout provides a continuous sediment supply by down-wearing its deflation basin. The dune may cease to increase in size if the funnelling effect is lessened by progressive widening of the openings, if a groundwater table or resistant stratum are exposed (such as a caliche bed or a clay bed), or if revigorated growth of vegetation impedes further erosion (Cooper, 1958; Livingstone and Warren, 1996; Melton, 1940). Baas and Nield (2007) simulated the development of blowouts using a Discrete ECo-geomorphic Aeolian Landscape model (DECAL), and showed the significant role of the interplay between vegetation growth and sand transport in the transformation from blowouts to elongated parabolic dunes. Baas and Nield (2010) further showed that blowouts can also expand laterally as they elongate, incorporate progressively increased sand, and develop into more mobile transgressive and transverse ridges. The initiation of blowouts by disturbances and their potential impacts on overall dunefield activation has been conceptualised by Barchyn and Hugenholtz (2013) in the context of vegetation resilience and sediment transport activity. They also showed that depth-limited blowouts can migrate and elongate at a high rate, and are hence more difficult to be stabilised by vegetation. In addition to cellular automaton models, Duran et al. (2008) applied a continuum model in a similar context to simulate the development of a parabolic dune in north-eastern Brazil.

The formation of parabolic dunes from blowouts requires the following three conditions: 1) a generally stabilised surface, which enables concentration of winds at isolated points of weakness; 2) an underlying sandy substratum, which provides sufficient sediment supply for forming a parabolic-shaped depositional lobe that must avoid the coalescence of adjacent blowouts; and 3) predominantly unidirectional winds (Cooper, 1958; Eriksson et al., 1989; Pye, 1983b). Blowouts developed on coastal dunes are in some instances regarded as a symptom of a negative sand budget (Livingstone and Warren, 1996; Psuty, 1988). A decrease in sand supply from a beach means that onshore winds have larger capacity to erode foredunes. The higher parts of the dunes are usually more vulnerable to desiccation and disturbance, and therefore more susceptible to initiation of a blowout. Blowouts are, however, present widely on the coasts of Manawatu Plain in New Zealand where beaches are progradational (Hesp, 2002). Sufficient sand supply and strong unidirectional winds are essential for the further transformation of blowouts into parabolic dunes, which may be secured by climatic changes such as a stronger wind regime, a more arid climate, and a lowered groundwater table, or by land degradation induced by anthropogenic perturbation such as grazing and human recreational activities.

Table 4. Studies on transformations from blowouts to parabolic dunes.

Reference	Study Region	Method	Transformation Cause
Melton, 1940	southern High Plains, US	aerial photographic interpretation	more arid climate, lowered ground-water surface
Hack, 1941	Navajo Country, Arizona, US	field surveys and photo-engravings	periglacial winds on sand covered by sparse vegetation
Brothers, 1954	Auckland, New Zealand	field observation	burning-off, animal tracks and topographic difference
Cooper, 1958	Oregon, US	description	windward slope stabilisation of saucer blowouts
Ranwell, 1958	Anglesey, Wales, UK	field surveys	natural forces
Pye, 1982	Cape Bedform and Cape Flattery, Queensland, Australia	aerial photographic interpretation	natural forces (fire, cyclones and lightning strikes) and local differences in vegetation cover or topography
Pye, 1983b	Northern Cape York Peninsula, Queensland, Australia	field surveys, aerial photograph interpretation and sediment analyses	increase in windiness or reduction in rainfall
Anton and Vincent, 1986	Jafurah Desert, Eastern Province, Saudi Arabia	field surveys and aerial photographic interpretation	preferable deflation of sand sheet areas
Eriksson et al., 1989	southern Kalahari Desert, South Africa	field surveys and aerial photographic interpretation	deflation of bare patches shaded by trees assisted by biological processes
Carter et al., 1990	west coast, Australia	description	trough blowouts evolve into parabolic dunes
Klijn, 1990	Younger Dunes, Netherlands	C14 and historical dating	climatic changes (sea level rise, and increased storm frequency and surges)
Hesp, 2001	Manawatu, New Zealand	aerial photographic interpretation	natural forces aided by human recreational activity
Hesp, 2002	-	description	high energy wind coasts
Gutierrezelorza et al., 2005	Tierra de Pinares, Spain	field observation	climate changes
Baas and Nield, 2007	-	DECAL model	Dynamic interplay between sedimentation balance and vegetation effectiveness
Duran et al., 2008	Ceara, Brazil	field surveys, QuickBird panchromatic satellite images interpretation and continuous modelling	vegetation growth rate defined by two parameters: the initial vegetation growth rate when no sand transport occurs and the maximum height a plant can reach
Hansen et al. 2009	Green Mountain Beach dune, Holland, southeast shore of Lake Michigan, US	field surveys	steering of winds in the deflation area
Girardi and Davis, 2010	Walking Dunes, New York, US	aerial photographic interpretation and previous literatures	dune-vegetation interactions

4.4 Transformations from parabolic dunes to other dune morphologies

Under certain conditions, parabolic dunes may be transformed into other dune morphologies. Parabolic dunes are commonly developed in well-vegetated landscapes and under conditions of a restricted sediment supply (Hack, 1941; Lancaster, 1995). If a new sediment supply becomes available or if vegetation cover decreases below a certain level, parabolic dunes may be transformed into more mobile dunes (Livingstone and Warren, 1996; McKee, 1966). Some studies have indicated that parabolic dunes can lose vegetation, and are activated and transformed into transverse dunes (Anton and Vincent, 1986; Hack, 1941; Hesp, 2001). Some regions exhibit a downwind transition continuum from parabolic dunes to barchan dunes and transverse dunes (Pye and Tsoar, 1990). Studies on these transformations are, nevertheless, few and often limited to anecdotal descriptions (Table 5). The biomorphic interactions and physical processes underlying the activation of parabolic dunes and their transformations into highly mobile barchan dunes or transverse dunes have not been investigated in detail, whereas these transformations may have significant implications on local land management and social-economic development. Parabolic dunes may also develop into other dune forms such as dome dunes (Anton and Vincent, 1986).

Table 5. Studies on transformations from parabolic dunes to other dune morphologies

Transformation Type	Reference	Study Region	Method	Transformation Cause
parabolic to transverse and barchan	Hack, 1941	Navajo Country, Arizona, US	field surveys and photo-engravings	vegetation destroyed by external sands from blowouts
	Anton and Vincent, 1986	Jafurah Desert, Eastern Province, Saudi Arabia	field surveys and aerial photographic interpretation	decline of vegetation density possibly due to change in water tables, natural vegetation succession and over-grazing
	Shulmeister and Lees, 1992	Groote Eylandt, Australia	thermo-luminescence dating	decline in vegetation cover because of decreased precipitation and/or increased aboriginal burning
	Hesp, 2001	Manawatu, New Zealand	aerial photographic interpretation	human activity (burning, grazing, introduction of exotic species and wetland modification)
	García-Hidalgo et al., 2002	Duero Basin, Spain	aerial photographic interpretation	noses of parabolic dunes are stopped by water and the arms continues to move forward
parabolic to dome	Anton and Vincent, 1986	Jafurah Desert, Eastern Province, Saudi Arabia	field surveys and aerial photographic interpretation	vegetation colonisation in the deflation hollow due to lower groundwater salinity
parabolic to longitudinal	Meurisse et al., 2005	northern shore, France	stratigraphy, ^{14}C dating and sedimentology	climatic modifications, agricultural practices and sea level rise

5. A conceptual framework for understanding parabolic dune transformations

Parabolic dunes are distributed widely across a large range of climatic gradients associated with a variety of floristic regions responding to different environmental stresses. Parabolic dunes are found on coasts, river valleys and lake shores as well as the margins of deserts and steppes, where eco-geomorphic interactions are often highly sensitive to climatic variability and closely influenced by socio-economic activities (Livingstone and Warren, 1996). This paper outlines a conceptual framework to assist in understanding the development of parabolic dunes and related transformations into other dune morphologies. In turn, this framework may serve as a useful tool to help predict possible landscape development scenarios in different aeolian environments under climate changes. Dune landscapes are governed primarily by wind regime, sand availability, and vegetation cover (Hack, 1941). The following three determinant factors are therefore chosen in this framework: 1) sand availability; 2) wind strength (wind variability is disregarded because parabolic dunes are mostly developed under a unidirectional wind regime); and 3) drought stress, the major control for the growth of vegetation in an aeolian environment. The degree to which drought stress influences vegetation growth varies depending on the characteristics of vegetation species, such as the resilience to drought and the capability of withstanding wind erosion and sand burial. Drought stress is generally determined by the regional climate, and encompasses the combined impacts of temperature, precipitation, and groundwater dynamics.

Two parts are included in the framework. The first part illustrates how increases and decreases in wind strength and/or increases and decreases in existing drought stress lead to the development of parabolic dunes evolved from blowouts (mobilisation), or from barchan dunes and transverse dunes (stabilisation), as shown in Figure 20. Each pane represents an eco-geomorphic system with a different level of sand availability, in which horizontal and vertical axes denote changes in drought stress and wind strength respectively. Five representative panes are shown here for simplification, with sand availability increasing from left-most (Figure 20a) to right-most (Figure 20e).

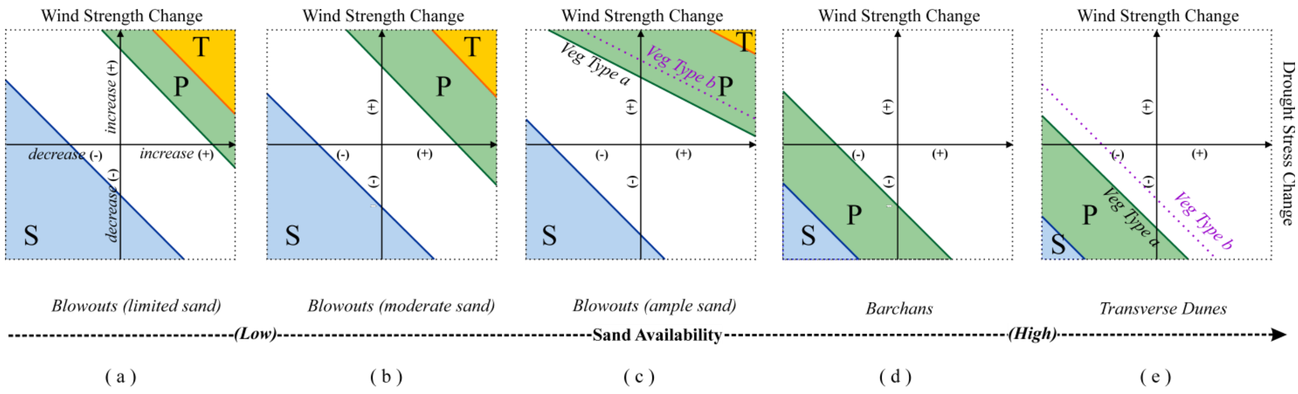

Figure 20. Conceptual diagrams of dune transformations leading to development of parabolic dunes under changes in wind strength and drought stress. Vertical axis within each pane denotes a change in wind strength, either increasing (upward) or decreasing (downward) from the original climatic state at the origin. Horizontal axis within each pane denotes a change in drought stress relative to the original climatic state, either increasing (to the right) or decreasing (to the left). S, P, and T represent Stabilised dunefields, Parabolic dunefields, and Transverse dunefields respectively. The white zones represent conditions where the existing landform is not transformed. Purple dotted lines are examples of minimal boundaries of changes in wind strength and drought stress leading to development of parabolic dunes under a different dominant vegetation species with stronger erosion and sand burial tolerances. The sequence of panes represents a gradient of sand availability from low (left-most) to high (right-most).

The transformation from a blowout to a parabolic dune (Figure 20a-c) occurs when sufficient sand is available from the deflation basin of the blowout for aeolian transport, and is deposited downwind, forming a tall, parabolic-shaped lobe.

In an environment with a limited sand supply from a sandy substratum (Figure 20a), given an existing vegetation species (with its specific growth structure and behaviour), a blowout is likely to be transformed into a parabolic dune if more sand is acquired by increasing wind strength and/or drought stress (moving to the upper-right in Figure 20a). Stronger winds have greater capability to erode soils and plants on the downwind edge of the blowout, and to expose more sand underneath as the blowout elongates by eliminating vegetation on its way. Greater drought stress provides more sand due to a loss of vegetation cover by desiccation. Extreme high wind strength may lead to the development of parabolic dunes from blowouts even if drought stress decreases slightly. Similarly, this transformation may happen during an extremely dry but less windy period.

We hypothesise that blowouts with a moderate sand supply (Figure 20b) can be transformed into parabolic dunes in an environment with smaller thresholds of increases in wind strength and/or drought stress compared with that of the blowouts in an environment with a limited sand supply. This difference may exist because more sand is available for transport when vegetation on the same surface area is removed.

In an environment with ample sand supply (Figure 20c), the available sand in a blowout is beyond the transport capacity of current winds, and wind strength becomes the limiting factor for sand transport. In this situation, an increase in drought stress alone (to any level) is insufficient to initiate a transformation, because sand transport is already at capacity. Drought stress impact is only effective if it is accompanied by an increase in wind strength capable of transporting the newly exposed sand. In a situation where drought stress does not change, an increase in wind strength leads more easily to a transformation to a parabolic dune than in the previous two cases (Figure 20a-b), because of a greater sand supply.

In all three cases discussed above, when wind strength and drought stress increase dramatically so that vegetation declines rapidly, blowouts elongate downwind, expand laterally, and interact/coalesce with nearby/adjacent blowouts, developing into a transverse dunefield. The thresholds of increased wind strength and/or drought stress needed for the transformation to transverse dunes are higher with greater sand availability (from Figure 20a to 20c): a thicker substratum below the blowouts offers a more abundant local sand supply, which causes the development of larger depositional lobes that migrate at a slower pace. Conversely, a thinner substratum enables blowouts to extend downwind at a higher rate and encourages the interactions and lateral calescence of nearby/adjacent blowouts. Large-scale coalescence activity and the disappearance of vegetated ridges between blowouts can then mobilise the entire area, leading to the development of a transverse dunefield (as discussed in section 4.3).

Blowouts can also be stabilised (moving to the lower left in Figure 20a-c) when wind strength decreases so that sand transport diminishes and the associated impacts of erosion and sand burial on vegetation growth are relieved, and/or drought stress decreases so that vegetation cover is restored due to greater water availability. Blowouts with less sand availability are more easily stabilised because of limited initial sand transport activity.

In comparison to blowouts (Figure 20a-c), barchan dunes and transverse dunes develop in an environment with a greater sand supply (Figure 20d-e). These dunefields are very active with a limited, usually temporary, vegetation cover.

The transformations of barchan dunes and transverse dunes into parabolic dunes (moving to the lower left in Figure 20d-e) are associated with the processes of dune re-vegetation and stabilisation, achieved by 1) a decrease in wind strength so that the erosion and sand burial activity is alleviated to the capability of vegetation tolerances; and/or 2) a decrease in drought stress so that less vegetation dies of desiccation. As more abundant sand is available in a transverse dunefield compared with that of a barchan dunefield, the transformation from transverse dunes to parabolic dunes requires a more considerable decrease in wind strength and/or drought stress in order to provide a relatively stable surface for vegetation taking root. An extreme situation in which both the wind strength and the drought stress are reduced significantly may lead to the stabilisation of the whole dunefield (moving to the lower left corner in Figure 20d-e).

Vegetation species plays a key role in dune transformations by moving the boundaries between different transformations with regard to changes in wind strength and drought stress.

An example in Figure 20c presents the boundary of a transformation from blowouts to parabolic dunes under the control of a different vegetation species with greater erosion and burial tolerances (*Veg Type b*), for example related to different growth structure and behaviour, in comparison to the default species (*Veg Type a*). To transform into a parabolic dune, a blowout keeps elongating downwind and eroding vegetation. A vegetation species with greater erosion and deposition tolerances is able to resist more severe sand transport activity. Therefore, the transformation from blowouts to parabolic dunes necessitates greater minimal increases in wind strength and drought stress.

The boundary of a transformation from transverse dunes to parabolic dunes regarding changes in wind strength and drought stress is more easily met (Figure 20e) under the influence of a vegetation species with greater erosion and sand burial tolerances (*Veg Type b*). This relation is due to the fact that stronger erosion and deposition tolerances enable vegetation to survive greater sand transport. As a result, smaller minimal decreases in wind strength and/or drought stress are required.

Similar deductions can be made for other dune transformations (e.g., from blowouts to transverse dunes, and from barchan dunes to parabolic dunes) and the stabilisation of different dunefields. In general, vegetation species with greater erosion and deposition tolerances move boundaries of dune transformations towards the upper right of panes in Figure 21.

The other half of the framework shown in Figure 21 illustrates how changes in wind strength and drought stress result in the transformations from stabilised as well as active parabolic dunes, with varying sand availability, into highly mobile barchan dunes and transverse dunes.

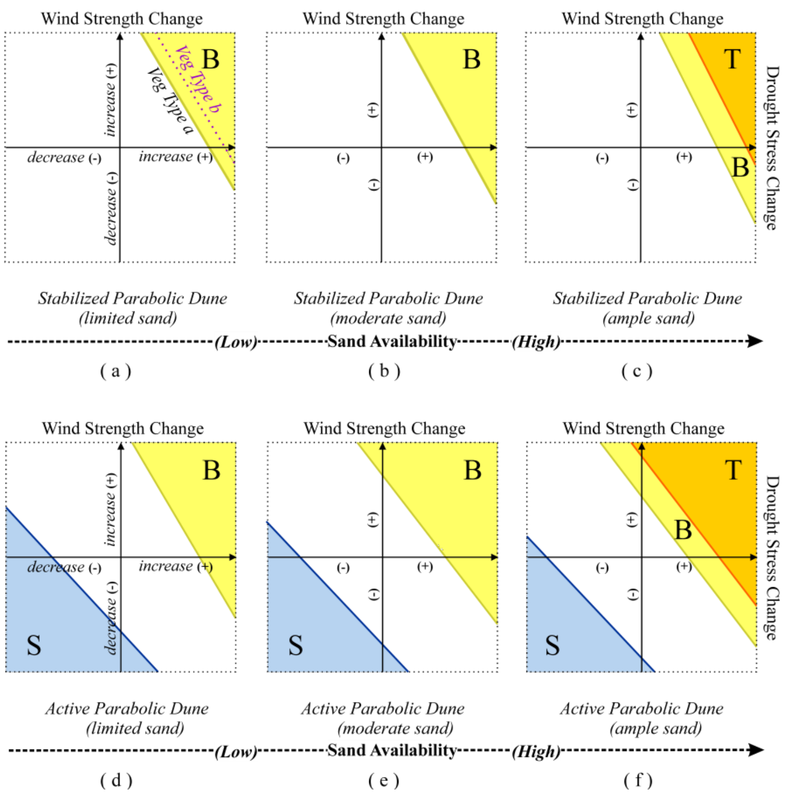

Figure 21. Conceptual diagrams of dune transformations from parabolic dunes to barchan dunes and transverse dunes under changes in wind strength and drought stress. Upper sequence represents transformations from fully stabilised parabolic dunes along a gradient of sand availability; lower sequence shows transformations from active parabolic dunes along a gradient of sand availability. Vertical axis within each pane denotes a change in wind strength, either increasing (upward) or decreasing (downward) from the original climatic state at the origin. Horizontal axis within each pane denotes a change in drought stress relative to the original climatic state, either increasing (to the right) or decreasing (to the left). B, T, and S represent Barchan dunefields, Transverse dunefields, and Stabilised dunefields respectively. The white zones represent conditions where the existing landform is not transformed. The purple dotted line in pane (a) is an example representing boundaries of changes in wind strength and drought stress leading to the activation of parabolic dunes into barchan dunes under a different dominant vegetation species with stronger erosion and sand burial tolerances.

Activation of stabilised parabolic dunes cannot occur solely by increasing wind strength, because surfaces are fully covered with vegetation and there is no sand available for transport (Figure 21a-c). A sole increase in drought stress, however, can lead to widespread vegetation decline due to desiccation, exposing sand at the surface to wind energy. Under severe drought stress, parabolic dunes are likely to be activated and transformed into barchan dunes even if wind strength decreases slightly. An increase in wind strength encourages the activation of parabolic dunes only if an accompanying drought stress makes available sufficient additional sand supply. The drought stress threshold for an increased wind strength being able to initiate a transformation from parabolic dunes to barchan dunes is independent of the sandy substratum thickness, but is governed by the characteristics of vegetation cover, in particular the vegetation resilience to drought and tolerances to erosion and sand burial.

Stabilised parabolic dunes with a thicker sandy substratum (from Figure 21a, to b, to c) require a smaller increase in drought stress to be transformed into barchan dunes, because more sand is exposed to winds under the same level of drought stress. When vegetation cover is largely destroyed by substantial increases in drought stress and/or wind strength, the activation of parabolic dunes can also lead to the development of transverse dunes, provided that sufficient sand is conserved in the dune environment (Figure 21c).

In comparison to stabilised parabolic dunes with the same level of sand availability (Figure 21a-c), the transformation of active parabolic dunes to barchan dunes requires relative smaller increases in wind strength and/or drought stress (Figure 21d-f).

For active parabolic dunes with a limited sand supply (Figure 21d), similar to their stabilised counterparts (Figure 21a), an increasing wind strength does not trigger a transformation until a greater drought stress creates an additional sand supply. The threshold of drought stress is lower for active parabolic dunes because more sand surface is already exposed to the wind. For the same reason, compared with that of their stabilised counterparts, the transformation from active parabolic dunes to barchan dunes requires a smaller increase in drought stress.

In a field of active parabolic dunes where sand availability is larger than sand transport capacity of winds (Figure 21e), parabolic dunes can be transformed into barchan dunes by increasing wind strength (and the associated sand transport capacity) alone, as well as by increasing drought stress. Although an increase in drought stress alone does not enhance sand transport capacity, an increase in drought stress raises the vulnerability of vegetation on trailing arms of active parabolic dunes and on surrounding interdune areas to wind erosion and sand burial activity. As a result, bare lobes of active parabolic dunes move forward more easily, then break with arms, and transform into barchan dunes.

In comparison to active parabolic dunes with a less sand supply, active parabolic dunes with an abundant sand supply (Figure 21f) require a smaller increase in wind strength and/or drought stress to be transformed into barchan dunes. A significant increase in wind strength and/or drought stress, meanwhile, can result in severe activation of dunes and the transformation of parabolic dunes into transverse dunes.

Active parabolic dunes can also be fully-stabilised when wind strength is reduced and/or drought stress is attenuated. Aeolian environments with more sand availability require larger decreases in wind strength and/or drought stress in order to be stabilised by vegetation (Figure 21d-f).

Similar to the dune transformations in Figure 20, the boundaries of transformations from active parabolic dunes to barchan dunes and transverse dunes as well as stabilisation by vegetation are controlled by the characteristics of vegetation species. An example in Figure 21a shows the boundary of the transformation from stabilised parabolic dunes with a limited sand supply into barchan dunes under the influence of a different vegetation species with greater erosion and sand burial tolerances (*Veg Type b*). In Figure 21, the boundary denoting thresholds for increases in wind strength and drought stress for this different species moves towards the upper right of the pane because vegetation is able to survive in an environment with greater aeolian activity. Similarly, boundaries of other transformations in Figure 21 show the same behaviour. Specifically, vegetation species with greater resilience to aeolian sediment mobilisation move transformation boundaries towards the upper right of a system in Figure 21.

As discussed in section 2, a change in dune migration rate is not a good indicator for the state of activity of an aeolian environment. Vegetation responds to climatic fluctuations generally more quickly than dune morphology, thereby providing a buffer between climatic change and geomorphic responses (Phillips, 1995; Yizhaq et al., 2009). Dune transformation tendency, representing the evolutionary trend of dune landforms mediated by vegetation change and eco-geomorphic feedback, is closely related to the stability of a dunefield and can be potentially used as an indicator to anticipate the activity of aeolian dunes under the impacts of environmental and anthropogenic forces.

Dune transformation tendency can be defined as the likelihood of a dune transformation occurring under changes in wind strength and drought stress, arising from environmental fluctuations, climatic change, and human disturbances. For example, blowouts with a limited sand supply need relatively large increases in wind strength and/or drought stress to transform into parabolic dunes, as compared to blowouts with a moderate sand supply, and therefore the transformation tendency from blowouts to parabolic dunes is relatively large in settings with a moderate sand supply. In other words, blowouts with a moderate sand supply are more sensitive to changes in environmental controls on their transformations into parabolic dunes. Dune transformation tendency is influenced by regional environmental parameters and can be potentially identified through knowledge of local eco-geomorphic dynamics and processes on a time scale of decades (Viles and Goudie, 2003).

The framework presented in this paper provides a conceptual way to organise and compare different aeolian dune environments and potential dune transformations under the impacts of environmental fluctuations and climatic change. The representative aeolian environments shown in the diagrams, however, are not fixed, and any given dune in reality can be related to different diagrams when sand supply changes. For example, sand availability can increase when more sand is blown from a beach into a field of parabolic dunes due to a rise in sea level (Carter, 1991). A fall of the groundwater table can also increase sand availability by increasing the thickness of a dry sandy substratum which can be potentially mobilised and transported by winds (Swezey, 2003). Furthermore, changes of the dominant vegetation species arising from either natural succession or human disturbances can move the boundaries of transformations in an aeolian dune system.

6. Conclusions

The development of parabolic aeolian dunes and related transformations to and from other dune morphologies are very sensitive to vegetation characteristics and to environmental variations arising from both natural and anthropogenic disturbances. This paper reviews the literature on parabolic dunes and their related transformations on a global scale to provide a comprehensive inventory. Coastal and inland parabolic dunes differ on environmental controls, transformation processes, and morphology. Coastal parabolic dunes are controlled by geometrical alignment of the coastline, tidal range, wave power, and sea level change. Coastal parabolic dunes are often associated with the initiation of blowouts on previously vegetated foredunes by either natural forces such as storms or human disturbances such as grazing, and are sometimes associated with the stabilisation of transgressive dunefields by vegetation colonisation. In contrast, inland parabolic dunes are governed by orographic conditions and groundwater availability related to nearby rivers or lakes. Inland parabolic dunes are largely transformed from barchan dunes or transverse dunes, which are usually found in arid and semi-arid regions. In relatively humid areas, coastal parabolic dunes can migrate over trees and form arms of relatively high relief, whereas inland parabolic dunes usually have arms of relatively low relief because grasses and shrubs dominate those areas. Elongated parabolic dunes are often found on coasts where wet periods alternate with dry periods accompanied by strong onshore winds (usually in equatorial or warm climates), but such elongated dunes are not commonly seen inland because the dune arms are frequently overridden or cut though by following dunes.

The transformation process from barchan dunes to parabolic dunes varies depending on vegetation species as well as the limiting factors of their germination and growth. The ‘horns-anchoring’ transformation mechanism is likely to happen when water deficiency is the limiting factor, whereas the ‘nebkhas-initiation’ transformation mechanism is likely to occur when vegetation growth is threatened primarily by wind erosion. Of the two mechanisms, the ‘nebkhas-initiation’ mechanism is less common because it requires more specialised vegetation species that can develop extensive roots utilising groundwater, while being able to withstand severe sand burial. A number of studies have explored the transformation from barchan dunes or blowouts to parabolic dunes. The transformations of parabolic dunes back into highly mobile barchan dunes or transverse dunes, however, have not drawn sufficient attention.

As parabolic dunes are widely distributed around the world across a broad climatic gradient, migration rates reported in literature vary because of local environmental settings, but also due to spatio-temporal scales and measuring methods employed. A change in the dune migration rate is, therefore, not a good indicator for the state of activity of an aeolian environment. The dune transformation tendency can potentially be used as a proxy for the stability of a dune system. The integrated framework presented in this paper provides a baseline for organising and comparing different aeolian dune environments, and for predicting possible transformation scenarios under changes of environmental controls. This framework shows how the impacts of changes in sand availability, wind strength, and drought stress vary in different aeolian settings. Further research is needed to quantify the exact thresholds of different transformations in various aeolian environments. Multiregional comparisons across the globe remain a challenge and need a wide collaboration among different research communities. The integration of real-world surveying with computer modelling may serve as a useful approach to further testing of this framework.

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